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MONTHLY WEATHER REVIEW

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REFRACTION PHENOMENA AFFECTING CEILOMETER OBSERVATIONS

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[Manuscript received March 6, 1950]

ABSTRACT

A typical ceilometer record during rainfall is examined and certain regular and recurring oscillations beginning uniformly at an angle of elevation of 48° and decreasing gradually with elevation are noted. These features are explained physically as "rainbow" effects. The angle of elevation of maximum oscillation is shown to be about 49° regardless of base-line length; the tapering off above is discussed. Possible "secondary rainbow" effects at lower elevation angles are mentioned. Finally, methods of differentiating between the rainbow effects and cloud layers are suggested.

INTRODUCTION

The standard Weather Bureau ceilometer usually gives an excellent record of the important meteorological element, cloud height [1]. However, with the onset of rain the record becomes confused by several peculiarities which are not easily explained at first glance. The chief characteristics of the record during a rainstorm, as indicated in figure 1, appear to be an oscillation which begins rather sharply at an angle of elevation of about 48° and which continues with decreasing intensity as the angle increases, and a second but intense oscillation at about the 80° elevation. These features of this particular record (fig. 1) are not unique, but are common to situations in which rain is falling from middle clouds in the absence of lower clouds. Variations are noticed in the height of the upper level at which the oscillations cease, but in nearly all cases the oscillation begins near 48°.

It has been noted that this "spot" just above the 48° elevation occurs at various stations throughout the United States.[2] The persistency with which this spot tends to occur at this angular elevation regardless of the base line would seem to indicate that the causes may be due to a raindrop refraction phenomenon rather than a diffused spot on a cloud base. In fact, since the ceilometer beam will not penetrate an appreciable thickness of cloud, it must be concluded that there are practically no clouds in the beam until it is extinguished at the highest level.

The aim of this paper is to study the ceilometer beam from a standpoint of atmospheric optics for the purpose of more accurately interpreting the record. More fully stated, the objectives of the paper are:

1. To demonstrate that when rain is falling, any spot showing in the ceilometer beam at (or very near) 49° should be regarded with caution in determining the height of the cloud base.
2. To present an explanation of other commonly observed oscillations in the ceilometer record during periods of rain. These consist mainly of:
   a. Attenuation of the spot with increasing elevation above 49°.
   b. Recurring oscillations near 36°.
   c. Recurring oscillations near the ground.
3. To develop criteria for deciding whether a spot is a reflection from a true cloud layer or merely a "rainbow" effect of the ceilometer beam shining through the raindrops.

PHYSICAL EXPLANATION OF SPOT NEAR 49°

The physical explanation of the refraction and reflection of sunlight by raindrops to cause the rainbow has long been known [2]. The light from the ceilometer projector

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[2] Mr. Ray Granger of the Oakland observing staff observed these spots at various stations. To him goes the credit for first pointing out such a spot to the writer and giving an indication of its true nature. Also as a direct result of Mr. Granger's alertness, the Weather Bureau Central Office issued special instructions to all ceilometer stations, pointing out the existence of this phenomenon and warning that it must not be confused with an actual cloud base. This phenomenon has been verified several times at the Air Force-Navy-Civil Landing Aids Experiment Station, Arcata, Calif., from simultaneous records by two ceilometers with different base lines (see L. A. E. S., *Final Reports, 1949*, Arcata, Calif., p. 27).

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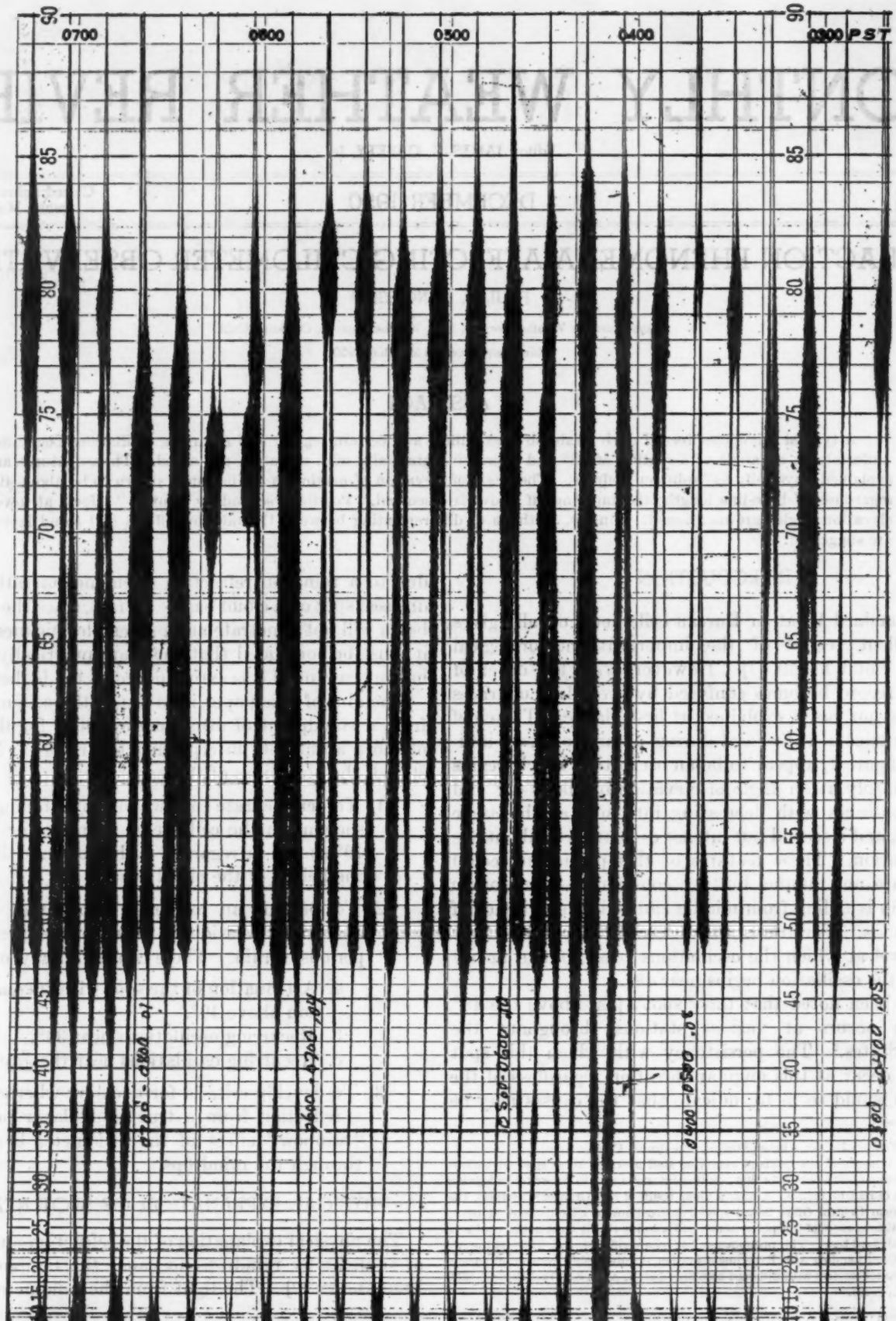

FIGURE 1.—A copy of a ceilometer record during a rain period. Oakland, Calif., December 18, 1949, 0300 to 0700 PST.

shining through rain presents a situation similar to that which results in the rainbow except that the beam is vertical and of very limited extent, and consists of but one color. In the following treatment, the problem of refraction by raindrops has been modified to take these factors into account.

SOLUTION OF THE REFRACTION PROBLEM

In the investigation of just how it is possible for a nearly monochromatic mercury vapor light refracted by raindrops aloft to produce bright areas within the beam, which result in the oscillations observed in the ceilometer record, three assumptions are made: (1) the freely falling drops are spherical; (2) the light from the mercury vapor projector beam is monochromatic with a value [3] of 5003Å; (3) the light rays from the ceilometer projector are parallel.

In figure 2, the ray from the mercury vapor light source S enters the drop at A at an arbitrary angle of incidence i, and is refracted at the angle of refraction r within the drop. The ray is then reflected at B by the upper spherical surface of the drop and then refracted again as it emerges at C at the angle of emergence e. The problem is to get an expression for the angle of deviation D, which is the total change of direction which the light ray suffers between entrance and emergence from the drop.

First, it will be useful to find from the geometry of figure 2 the relation of r to r' and of i to e. Angle OBA equals angle OBC because the radius OB is also the axis of reflection of the spherical surface in the region of B. Hence, the two isosceles triangles AOB and BOC are congruent and $r'=r$. Now, since $\sin i=\mu \sin r$, and $\sin e=\mu \sin r'$ (where μ is the index of refraction), $i=e$; the angle of incidence equals the angle of emergence.

With these results, an expression for D, the total deviation of the light ray, may be found in terms of i and r. In going from S to A to B, the ray is deviated clockwise through an angle $(i-r)$. Since angle $ABC=2r$ (see above), in traversing the inside of the drop from A to B to C the ray is rotated clockwise about B through an angle $(180°-2r)$. Finally it is rotated clockwise through the angle $(e-r')$ in going from B to C to E. Summing these rotations gives a total deviation $D=(i-r)+(180°-2r)+(e-r')$. Since $(e-r')=(i-r)$,

$$D=180°+2i-4r. \qquad (1)$$

If parallel rays are incident on a sphere, the angles of incidence vary between 0° (for the ray traversing the center) and 90° (for the rays tangential at the edges). For any arbitrary angles of incidence i, corresponding values of r may be computed and substituted in equation (1) to determine D, the final deviation. A series of computations using different values for angle i yields the curve of D against i given in figure 3. This curve was computed for index of refraction $\mu=1.336$ when wavelength $\lambda=5003$Å, temperature $T=20°$ C., and pressure $p=1$ atmosphere (see [3]).

To find an expression for the minimum point of the curve, differentiate equation (1) and set $dD=0$. This gives $di=2dr$; but $\sin i=\mu \sin r$, and $\cos i\cdot di=\mu \cos r\cdot dr$. Thus, $\mu \cos r=2 \cos i$, or

$$\cos i=\sqrt{\frac{\mu^2-1}{3}} \qquad (2)$$

when D is a minimum. Setting $\mu=1.336$ in (2) gives $\cos i=0.5116$ and $i=59°14'$.

For this value of i, the angle of refraction equals 40°2′. Substituting $i=59°14'$ and $r=40°2'$ into (1) gives a minimum value of D equal to 138° 20′ for the beam from the

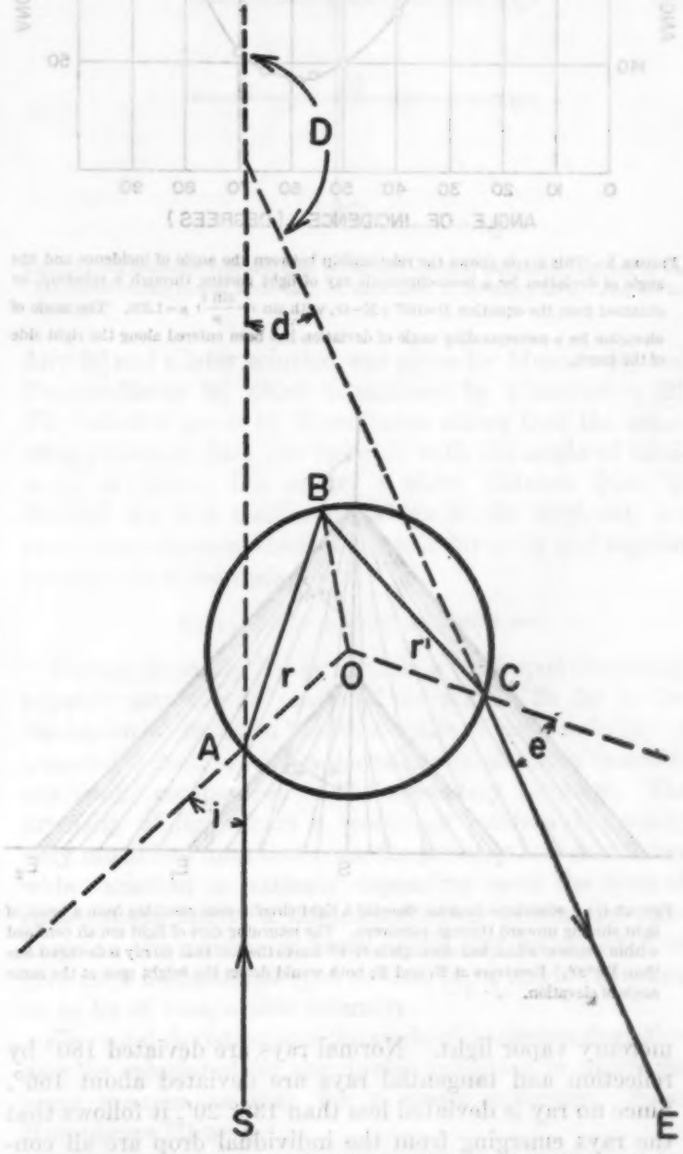

FIGURE 2.—Diagram showing the path which a vertical ray of light takes in passing through a raindrop near minimum deviation in which one internal reflection takes place. As shown by this diagram the angle of elevation equals $90°-d=D-90°$ (where $D+d=180°$).

[3] The mean value of the two principal mercury lines at wavelengths 4358Å and 5461Å is 4910Å; the nearest wavelength λ for which Dorsey [3] gives a good value of the index of refraction μ is $\lambda=5003$Å.

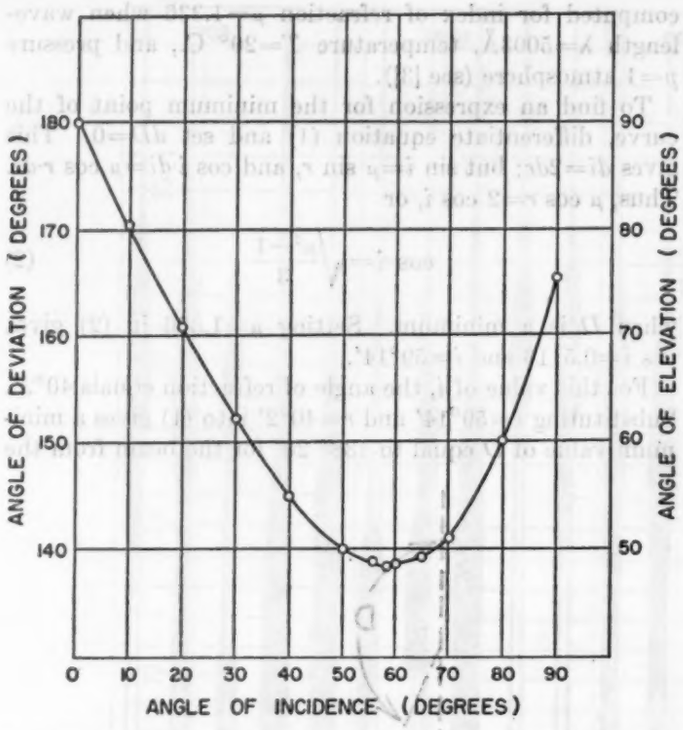

FIGURE 3.—This graph shows the relationship between the angle of incidence and the angle of deviation for a monochromatic ray of light passing through a raindrop, as obtained from the equation $D=180°+2i-4r$, with $\sin r=\frac{\sin i}{\mu}$, $\mu=1.336$. The angle of elevation for a corresponding angle of deviation has been entered along the right side of the graph.

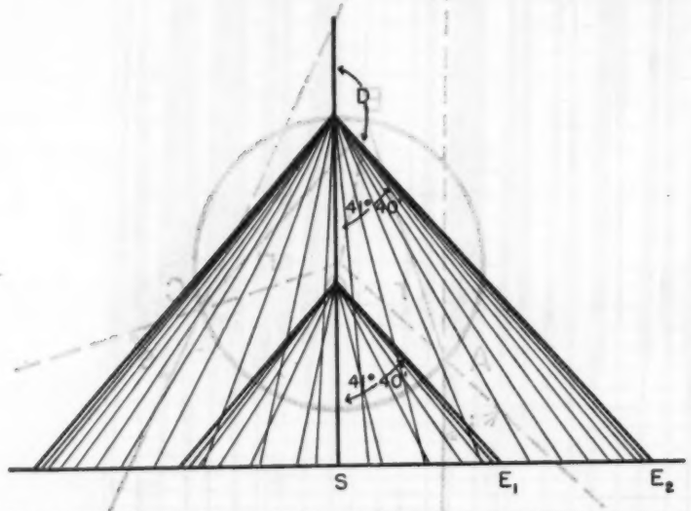

FIGURE 4.—A schematic diagram showing a right circular cone resulting from a beam of light shining upward through raindrops. The returning rays of light are all confined within a cone of which half the angle is 41°40′ due to the fact that no ray is deviated less than 138°20′. Receivers at E_1 and E_2 both would detect the bright spot at the same angle of elevation.

mercury vapor light. Normal rays are deviated 180° by reflection and tangential rays are deviated about 166°. Since no ray is deviated less than 138° 20′, it follows that the rays emerging from the individual drop are all contained within a right circular cone, of which half the vertical angle, d, equals 41° 40′ as illustrated in figure 4.

FORMATION OF SPOT NEAR 49°

Near the point of minimum deviation, as shown by figure 3, a relatively large change in the angle of incidence results in only a small change in the angle of deviation, D. Consequently, there is a tendency for light rays to be congested near the minimum angle of deviation of 138°20′, which corresponds to the angle of elevation of 48°20′ (the elevation angle$=D-90°=90°-d$). However, the greatest intensity of illumination in the beam occurs at a slightly larger deviation angle [2], thus causing the brightest area to appear just above the angle of elevation of 48°20′.

While the curve in figure 3 is constructed for a monochromatic light beam with an index of refraction of 1.336, a similar curve could be constructed for every individual wave length in the spectrum, the index of refraction, μ, varying with the wave length. For red light $\lambda=7065$Å, $\mu=1.330$; for violet light $\lambda=4102$Å, $\mu=1.342$. If curves like the above are constructed for these wave lengths, half of the vertical angle, d, of the cone of emergent rays is 42.1° for red light and 40.2° for violet. If the ceilometer beam were made up of all wave lengths, a very short section of a rainbow would be visible at the spot between 47.9° and 49.8° elevation with the red at the lower and the violet at the upper position. From this it may be seen that the spot observed at about the 49° elevation during rainfall may be considered as a short section of a "monochromatic rainbow" lying in a horizontal plane above the observer.

DEPARTURES FROM THEORY

It should be understood that the value of $d=41°40'$ given above is only a theoretical value dependent upon the initial assumptions. However, we would expect to find the observed value very near that indicated by theory as only small errors would be introduced into the value of d by departures from the initial assumptions. Slight variations in the value of d, with corresponding changes in the angle of elevation, would likely be caused by the following factors: (1) the size of the spot would be increased slightly by the dispersive effects of the nonspherical drops; (2) μ for distilled water varies slightly with the temperature (for $\lambda=5016$Å, $\mu_{10°C}=1.337070$, $\mu_{20°C}=1.3363453$, $\mu_{30°C}=1.335289$, a variation in the fourth figure only); (3) the combination of the two wave lengths 4358Å and 5461Å will result in more spreading of the spot than would be present in purely monochromatic light; (4) the lack of strict parallelism of the rays would introduce only a slight variation in d; (5) slight errors may be present in the ceilometer, thus affecting the measurement of the angle of elevation.

The first four (physical) factors, acting accumulatively, should not give a departure of more than one degree from the theoretical value. The fifth (instrumental) effect is purely a local problem. A consistent occurrence of these

spots at 45°, for example, should lead one to investigate whether or not the light has been jarred out of normal alignment.

PHYSICAL EXPLANATION OF OTHER FEATURES

The discussion thus far offers a satisfactory explanation for the formation of the main spot in the ceilometer beam at about 49°. Additional features noted on the ceilometer record and shown schematically in figure 5 remain to be explained. These features may be listed as follows: (1) The gradual weakening of the intensity of the oscillation between 49° and about 76°; (2) The apparent tendency for a weak spot to form at about 36°; (3) The existence of the oscillation near the surface, which occasionally extends upward until it merges with the spot at 49°.

DECREASE IN INTENSITY ABOVE 48°

The diminution of the oscillation with increasing angle above the main spot, *A*, as indicated in figure 5, appears to be nearly inversely proportional to the angle of elevation. A qualitative explanation of the decrease in intensity may be obtained through a consideration of the effectiveness of the raindrop in returning the light ray for varying angles of incidence and the increasing loss of light by scattering for the longer path lengths.

Each raindrop in the light beam deflects light to the receiver either by internal or external reflection, or both. Only those drops from the angle of elevation of 48°20′ upward to the cloud base will return light by one internal reflection in addition to that reflected at the surface. As previously mentioned, and as indicated in figure 3, there is a packing of the internally reflected rays near the angle of minimum deviation. For angles of deviation greater than this minimum value, but less than 166° (see fig. 3), there are two rays with different angles of incidence which have the same deviation angle. For example, rays incident at angles of about 20° and 87° suffer the same total deviation. As the steepness of both arms of the curve increases, the congestion of the light rays diminishes for angles of incidence farther and farther away from the value of 59°14′. As a consequence, we should expect the light from the beam to decrease in intensity with elevation above the main spot.

The greater path lengths for portions of the beam at the higher elevations would also tend to diminish the amount of light returned, due to the increased scattering. Other factors, such as the variation of the amount of light reflected for varying angles of incidence, would have to be considered to obtain a complete solution to the problem. The decrease in the intensity of the light for angles of deviation greater than the minimum value has its counterpart in the well-known fact that the area beneath the rainbow is brighter than that above it.

A classical solution to the problem of the intensity of the wave front and its variation with angular distance from the ray of minimum deviation was first obtained by

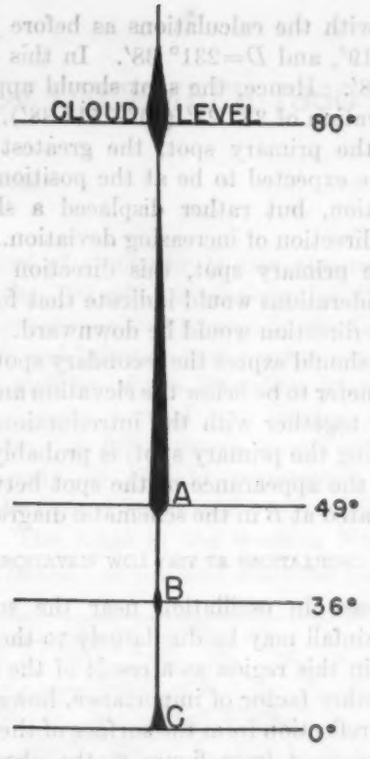

FIGURE 5.—A schematic representation of the most common oscillations recorded by the ceilometer during periods of rainfall. *A*, main spot; *B*, secondary spot; *C*, surface fog and haze.

Airy [4] and a later solution was given by Mascart [5] and Pernter-Exner [6] which is outlined by Humphreys [2]. The solution given by Humphreys shows that the maximum intensity does not coincide with the angle of minimum deviation, but occurs a short distance from it. Beyond the first maximum, which is the brightest, are succeeding maxima which decrease in intensity and angular intervals at a decreasing rate.

FORMATION OF A WEAK SPOT NEAR 36°

During periods of heavy rainfall, a weak spot frequently appears near the 36° angle of elevation. So far in the discussion no mention has been made of the possibility of a second reflection within the raindrop causing a "secondary spot" comparable to the secondary rainbow. The intensity of light from a secondary rainbow is usually very much less than that from the primary bow and shows wide variation in intensity depending upon the type of rain. Accordingly, a double reflection within the raindrops in the ceilometer beam would not be expected to give such a consistent spot as that for a single reflection or to be of comparable intensity.

The total deviation and the angle of minimum deviation may be calculated for the double reflection in much the same manner as that for a single reflection. From Humphreys [2] we get

$$D=360°+2i-6r \text{ and } \cos i=\sqrt{\frac{\mu^2-1}{8}}$$

Proceeding with the calculations as before yields $i=71°\ 46'$, $r=45°\ 19'$, and $D=231°\ 38'$. In this case, $d=D-180°=51°\ 38'$. Hence, the spot should appear near the angle of elevation of $38°\ 22'$ ($90°-51°\ 38'$).

As with the primary spot, the greatest illumination would not be expected to be at the position of the minimum deviation, but rather displaced a short distance toward the direction of increasing deviation. It was seen that for the primary spot, this direction was upward; similar considerations would indicate that for the secondary spot the direction would be downward. Thus, qualitatively, we should expect the secondary spot as picked up by the ceilometer to be below the elevation angle of $38°\ 22'$. This factor, together with the introduction of the same errors affecting the primary spot, is probably sufficient to account for the appearance of the spot between 36° and 37°, as indicated at *B* in the schematic diagram in figure 5.

OSCILLATIONS AT VERY LOW ELEVATIONS

The increase in oscillation near the surface during periods of rainfall may be due largely to the formation of fog or haze in this region as a result of the precipitation. There is another factor of importance, however, and that is the direct reflection from the surface of the raindrops.

As may be seen from figure 6, the observer receives reflected light from a whole column of drops in the light beam, the angle of reflection decreasing with the height of the drop. The greatest amount of reflected light would be from a ray reflected tangential to the surface of a drop infinitely far below the observer. Since the observer is not able to see drops lower than the ceiling projector, the amount of reflected light would be greatest for a 0° angle of elevation and would decrease with an increasing angle.

During periods of heavy rain a continuous oscillation is occasionally recorded between *B* and *C* in figure 5. This type of oscillation is probably caused to a large extent by the external reflection from the drops. The more common oscillation occurring between 0° and 15° elevation (see fig. 5) more likely results from haze and fog forming in this region.

PRACTICAL APPLICATIONS

Now that the various ways in which the ceiling light and rain drops can produce oscillations in the receiver have been investigated, means of differentiating between refraction phenomena and actual cloud layers will be discussed.

At night the problem is simplified by the fact that the spot is visible. All the observer need do is walk a short distance toward the ceiling light; if the elevation angle of the spot remains about 49°, the light is not being reflected from a cloud layer.

In the daytime other methods must be used. If the following points are kept in mind, the observer should be able to recognize the refraction phenomena from the ceilogram alone; (1) the spots occur regularly near 49° elevation as long as rain continues; (2) the intensity of the oscillation should follow roughly the intensity of rainfall; (3) all of the refraction phenomena described allow further penetration of the beam until a true cloud layer is reached at a higher elevation (e. g., at about 80° as shown in fig. 5). Ceiling balloons and frequent checks with incoming and outgoing aircraft may be used if doubt still exists.

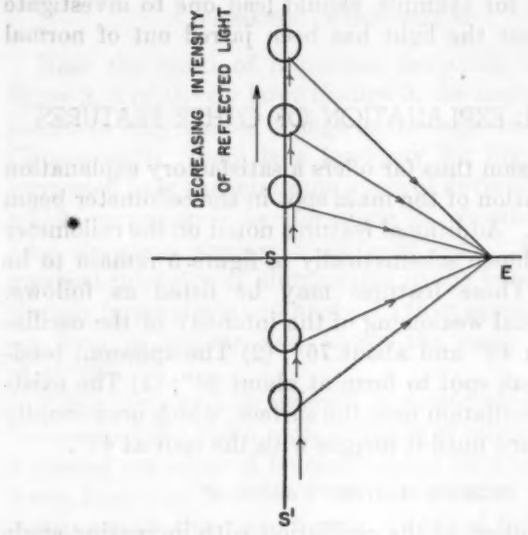

FIGURE 6.—Diagram indicating the decreasing angle of incidence with increasing elevation angle of the raindrop; the decreasing angle of incidence results in a diminishing of the light returned to the receiver at *E*.

ACKNOWLEDGMENT

The writer wishes to express his gratitude for the valuable help rendered by Mr. Donald L. Jorgensen in preparing the manuscript. Mr. Jorgensen also pointed out the importance of the gradually decreasing oscillations above 49° and the probable cause.

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THE WEATHER AND CIRCULATION OF DECEMBER 1950[1]

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The weather and circulation of December 1950 are notable for the degree of persistence which operated from the preceding month. The monthly temperature and precipitation anomalies in the United States for the two periods are remarkably similar. The mean temperatures in the West persisted above the seasonal normals, while in the East, with the exception of New England, the weather was colder than normal. Precipitation continued abundant in New England, the Great Lakes region, and the Northwest, while drought conditions prevailed over the Southern Plains and the Southwest.

A great deal of similarity is also apparent in several features of the 700-mb. circulations for the 2 months (fig. 1). A low center was present in the Gulf of Alaska with a trough and below normal heights extending southward to the vicinity of the Hawaiian Islands. A ridge with heights above normal was located over western North America. A full latitude trough extending from northern Canada to Florida maintained its greatest intensity in the United States and was weaker than normal in Canada. The ridge in the western North Atlantic,

[1] See charts I–XI following p. 225, for analyzed climatological data for the month.

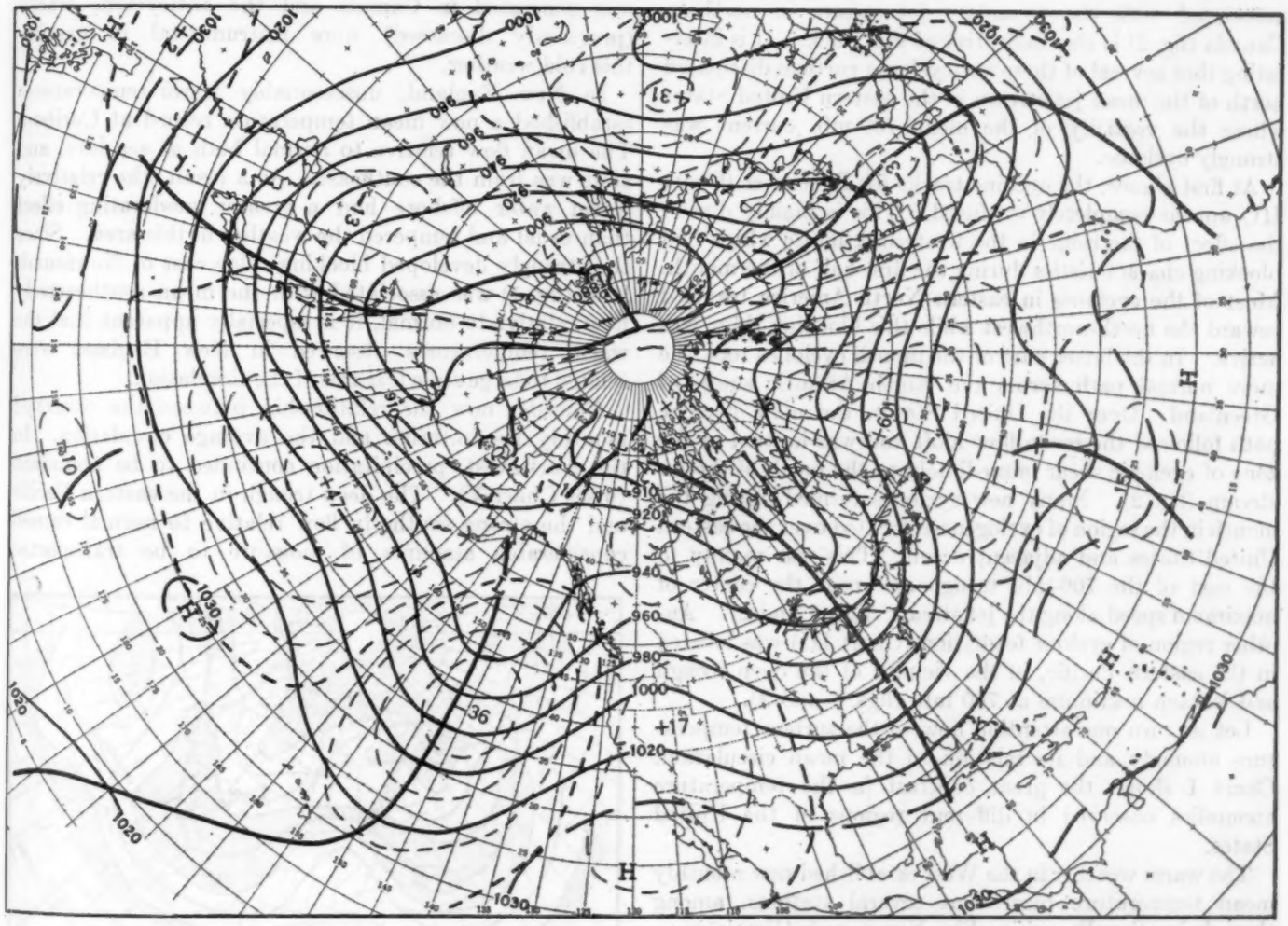

FIGURE 1.—Mean 700-mb. chart for the 30-day period November 28–December 27, 1950. Contours at 200-ft. intervals are shown by solid lines, 700-mb. height departures from normal at 100-ft. intervals by dashed lines with the zero isopleth heavier. Anomaly centers and contours are labeled in tens of feet. Minimum latitude trough locations are shown by heavy solid lines.

however, had more pronounced blocking characteristics in December when a well defined split in the 700-mb. jet stream was apparent (fig. 2). Other features of the jet stream are similar to those of the preceding month with the exception of the greater amplitude to the meanders over the United States. It also appears significant that the center of maximum wind speed which was located upstream from the weakly reverse-tilted (NNW–SSE) trough in the eastern United States during November was found slightly to the east of the north-south oriented trough in December.

The anticyclone tracks over North America (Chart II) continued from northwestern Canada through the Northern Plains to the Southeast. In northwest Canada, where these anticyclones originated, a mean sea level High was located with central pressure six millibars above normal. The elongation of this sea level anticyclone southeastward into the Northern Plains (not shown) very clearly defines the main anticyclone track. The anticyclonic vorticity associated with the secondary jet stream in northern Canada (fig. 2) is also indicative of this track. It is interesting that several of these anticyclonic vortices dissipated north of the mean jet stream in the eastern United States where the vorticity of the mean 700-mb. current was strongly cyclonic.

At first glance, the cyclone tracks for December (Chart III) appear completely confused. This is mainly due to the effect of the ridge in the western Atlantic which had blocking characteristics during the first half of the month. Most of the cyclones in eastern North America traveled toward the north-northwest while this blocking ridge was active. In the latter part of the month cyclones took the more normal path across the North Atlantic south of Greenland. Over the United States, the main cyclone path followed the mean flow aloft and was located in the zone of cyclonic shear immediately to the north of the jet stream (fig. 2). Many new cyclones formed during the month in the region of cyclogenesis located over the eastern United States and adjacent ocean. This was mainly to the east of the 700-mb. trough and near the center of maximum speed along the jet stream (figs. 1 and 2). Another region of cyclone formation (Chart III) was located in the eastern Pacific, in the vicinity of the deep trough and isotach maximum at 700 mb. (figs. 1 and 2).

Let us turn our attention now to the surface temperature anomaly and its relation to the mean circulation. Chart I shows the great contrast in the temperature anomalies observed in different regions of the United States.

The warm weather in the West established new monthly mean temperature records at several stations, among these being San Francisco, Ely, Yuma, and Albuquerque. The mean 700-mb. circulation (fig. 1) shows that the air entering the West had a more southerly trajectory than normal. This can be seen from the meridional flow indicated by the height anomaly of the 700-mb. surface. The above-normal heights in the West are also related to the warm temperatures. In the Great Basin, where a sea level anticyclone was observed (Chart VI), the unusual lack of snowcover (inset Chart VII) contributed to the warm weather. In December, a Great Basin High is frequently associated with cool mean temperatures, but the lack of snowcover in this case allowed less radiative cooling than normal. The persistence of valley fogs in Utah and Nevada for approximately two-thirds of the month was reflected in the low percentage of clear sky observed (Chart IV), and also reduced the radiative cooling of the surface.

Below normal temperatures were observed from the Great Lakes to Florida and several new daily records were set. The 700-mb. heights were below normal over this region and the flow aloft was northerly relative to normal. Several cold waves accompanied the polar continental anticyclones which entered this region from the northwest. The unusually intense polar air which was generated in Canada and the anticyclone tracks (previously discussed) were instrumental in causing this cold weather.

In New England, unseasonably warm temperatures established a new mean temperature record at Caribou. The mean flow relative to normal both at sea level and aloft was from the southeast. As a result, the relatively warm water offshore had a greater moderating effect than usual and tempered the weather in this area. Since the strongly developed blocking ridge east of Newfoundland (fig. 1) was associated with the mean southeasterly flow relative-to-normal, it is especially apparent that the warm temperatures observed in New England were related to large scale features of the circulation.

Consider now the relationship between the observed monthly precipitation and the average circulation. In the Northwest, precipitation continued to be abundant (inset Chart V). The deep trough in the eastern Pacific and the strong southerly flow relative to normal caused considerable amounts of moisture to be transported

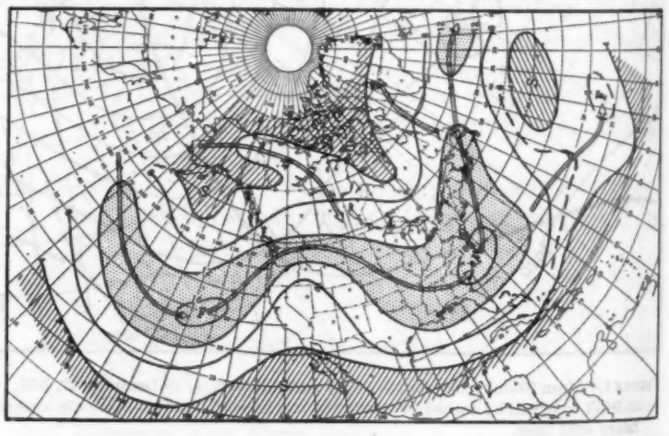

FIGURE 2.—Mean geostrophic (total horizontal) wind speed at 700-mb. for the 30-day period November 28–December 27, 1950. Isotachs at 4-m. p. s. intervals are shown by solid lines while the axes of maximum wind speeds (jets) are shown by double lines.

northward into the westerlies along the West Coast. In addition, orographic lifting, which is related to the strength of the westerlies in this area, was slightly greater than normal. The precipitation in this area would have been heavier except for the mean anticyclonic vorticity and the positive anomaly of 700-mb. heights both of which restricted the cyclonic activity (Chart III).

A deficiency of precipitation persisted from November in the middle and southern Great Plains and the Southwest. Drought conditions in the southern Great Plains caused poor growth of winter wheat and pastures deteriorated over most of the region. A new monthly record for deficient precipitation was set at Fort Worth while the records were tied at El Paso and Presidio, Tex., and Albuquerque, N. Mex. The flow aloft was from the northwest and northerly relative to normal. This decreased the usual supply of moisture from the Gulf of Mexico. The air entering this region was mainly of Pacific origin and extensive foehn drying took place. The cyclonic activity in this area was greatly restricted (Chart III) due to the mean anticyclonic vorticity associated with the jet stream to the north and the strong ridge in the west.

In the vicinity of the Great Lakes and the Northeast, copious amounts of precipitation were recorded for this month mainly in the form of snow (see Chart VII). Minneapolis reported fifteen consecutive days of snow and Duluth had the excessive amount of 25 inches in a 24-hour period. The excessive precipitation in the region surrounding Duluth was related to the easterly flow relative to normal at sea level and aloft accompanied by the warming and moisture source of the Great Lakes. A deep mean trough aloft was centered in Ohio and the associated mean cyclonic vorticity extended both to the west and east.

Normal amounts of precipitation were recorded in the Southeast. The location and intensity of the mean trough were favorable for precipitation in this region. Greater precipitation would have been expected if the trough had a northeast-southwest tilt.

A DECEMBER STORM ACCOMPANIED BY TORNADOES

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INTRODUCTION

On December 2, 1950, a storm over the central United States was accompanied by three tornadoes in Illinois and one in Arkansas. In Illinois tornadoes are relatively rare during December, a total of only 3 days with tornadoes, one each in 1876, 1949, and 1950, and a total of six tornadoes, having been reported since 1835. Table 1 shows the approximate time of the tornadoes on December 2, 1950, their path, and extent of damage. These storms occurred in an elongated low pressure system which extended from northern Texas and Arkansas thence north-northeastward over Lake Michigan, and they were associated with polar maritime air aloft from the Pacific which had moved rapidly across the Plateau on November 30 and December 1.

It is difficult if not impossible in synoptic studies to point out the conditions that are unique to tornado formation because of the relatively small size of these storms as compared to the density of observational data. For this reason the following discussion will attempt only to point out some of the synoptic features frequently observed with tornadoes and which are evident in the situation of December 2.

PRECEDING SYNOPTIC CONDITIONS

The synoptic situation on the early morning of the preceding day is shown in figure 1, the surface map for 0630 GMT December 1. Significant features include the quasi-stationary front extending east-west across the Southern States and the maritime Pacific cold front moving eastward through Idaho, Nevada, and California. At the same time, gradient level winds over Louisiana and eastern Texas were southerly at about 40 knots; they had increased in speed and were bringing warm moist air over the central United States. Along with this northward flow of warm air, the front across the Southern States began thereafter to move rapidly northward as a warm front. Twenty-four hours later (fig. 2) this front had moved into Illinois and Indiana, while the Pacific cold front extended through Texas, Oklahoma, and aloft over Kansas. Northward flow of warm moist air had further increased, so that maritime tropical air with its unseasonably high surface temperature and moisture content extended into southern Illinois.

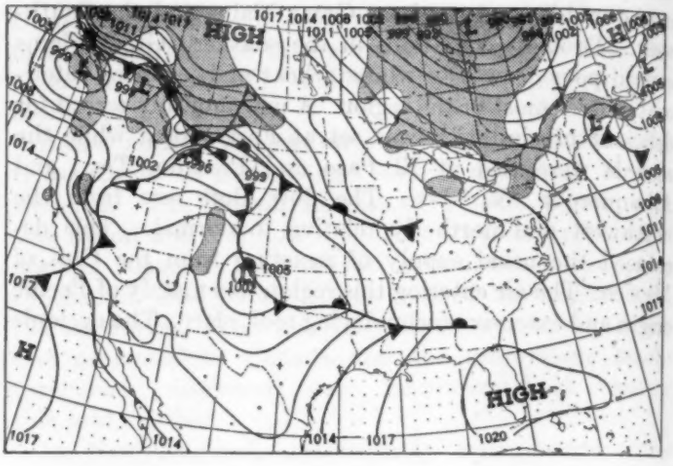

FIGURE 1.—Surface weather chart for 0630 GMT, December 1, 1950. Shading indicates areas of active precipitation.

Progress of the leading edge of polar maritime air from the Pacific can be seen in figure 3, where positions of the Pacific front are shown at 3-hour intervals. Beginning near or shortly after 0630 GMT of the 2d, this front was no longer evident at the surface. This means that the

TABLE 1.—*Tornadoes on Dec. 2, 1950*

Location of origin	Approximate time of origin	Path	Location of termination	Approximate time of termination	Effects
ILLINOIS	*GMT*			*GMT*	
Near Fosterburg (Madison County)	2100	25 mi. ENE.	Near Mount Olive (Macoupin County)	2140	Slight damage, 1 death, 3 injuries.
Near Highland (Madison County)	2200	22 mi. ENE.	Greenville (Bond County)	2210	$500,000 damage, 2 deaths, 24 injuries.
4 mi. NW of Sparta (Randolph County)	2330	11 mi. ENE.			$45,000 damage, no deaths or injuries.
ARKANSAS					
Near La Cross (Izard County)	2225	8 mi. NE.	Near Myron (Izard County)		Slight damage, no deaths or injuries.

NOTE: Since time of occurrence and location are approximate, the data given will not accurately represent the speed of the tornadoes.

leading portion of the polar maritime air was overriding the surface layer of maritime tropical air, a condition that is frequently observed with tornado situations in the central United States. In such cases the polar maritime air, while relatively cold when passing over the Plateau, becomes progressively warmer in its movement over the mountain region and the higher portion of the east slope so that its leading edge acquires a potential temperature higher than that of the low level maritime tropical air over the central Plains. This is in agreement with the mechanism for tornado formation proposed by Lloyd [1] and is similar to a situation described by Fulks [2].

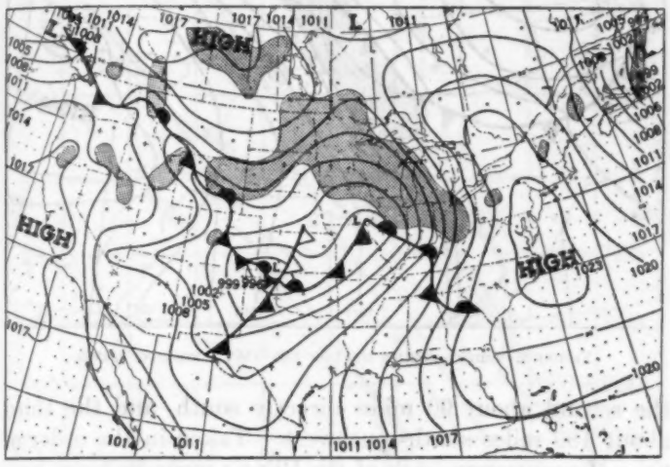

FIGURE 2.—Surface weather chart for 0630 GMT, December 2, 1950.

SYNOPTIC FEATURES DURING TORNADO CONDITIONS

Progress of the frontal systems during the time of tornado occurrences may be seen in figure 3. Figures 4, 5, and 6 show details of the synoptic situation at 1830 and 2130 GMT on December 2 and 0030 GMT December 3. The first tornado began near Fosterburg, Ill. (northernmost of small solid squares in fig. 5) about the time of figure 5, and the last (southernmost of the three tornadoes in Illinois) began near Sparta within the hour preceding the time of figure 6. The position of the main surface front and minor wave in Missouri and central Illinois as shown in figures 4, 5, and 6, is located to a fairly high degree of accuracy on the basis of hourly reports from St. Louis, Mo., Belleville, Rantoul, and Springfield, Ill., and other surounding stations; also a radar line-echo report from Belleville which was taken to be the cold front. The upper cold front is confirmed by upper air reports, particularly in six-hourly radiosonde ascents at Rantoul in eastern Illinois, but it cannot be located with the same precision as the surface front.

The evidence indicates with reasonable certainty that the tornadoes occurred between the surface cold front and the upper front, that is, in the area where tropical air was being overrun by polar maritime air. The distance ahead

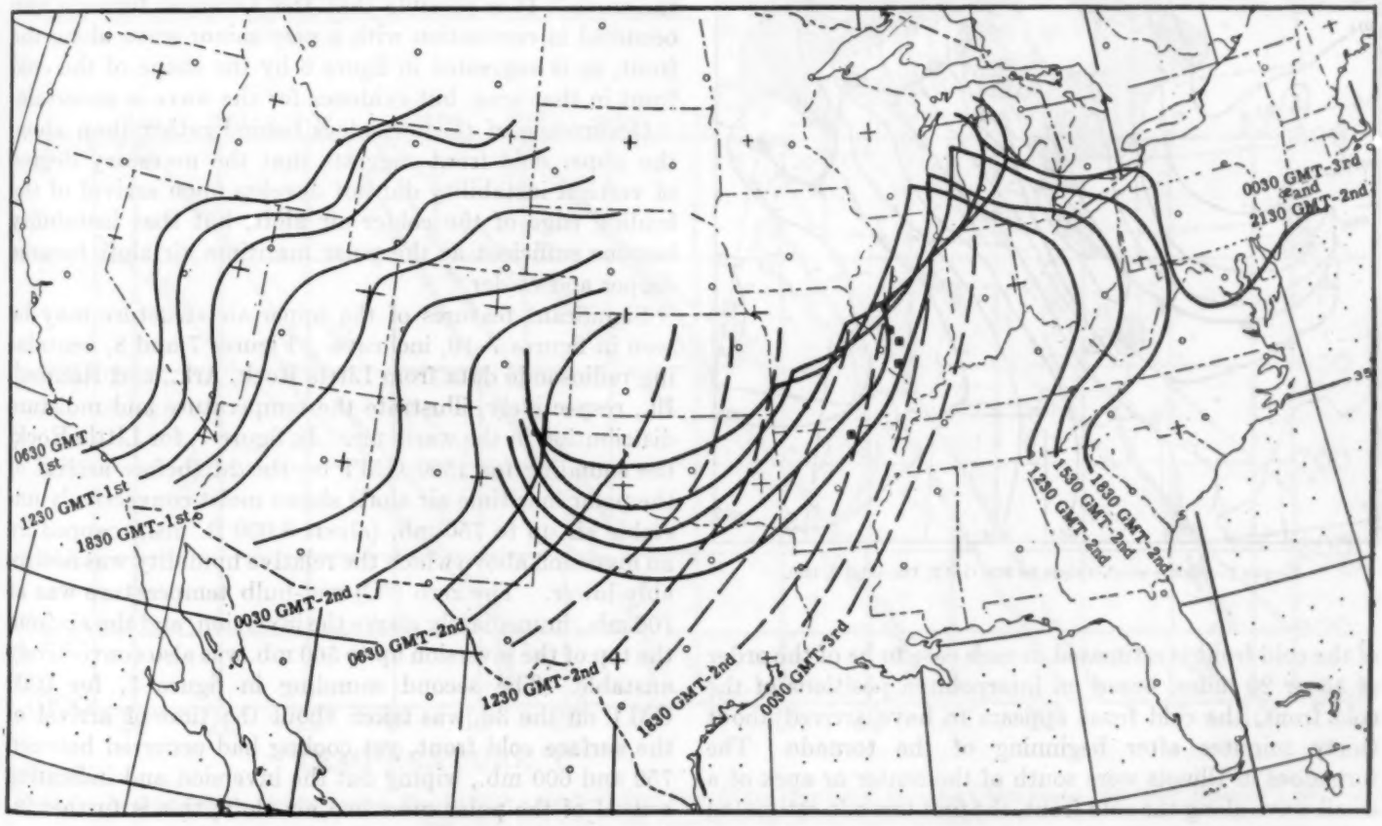

FIGURE 3.—Positions of the significant fronts at intervals of 3 hours. Tornadoes occurred between 2100 and 2400 GMT at locations indicated by black squares.

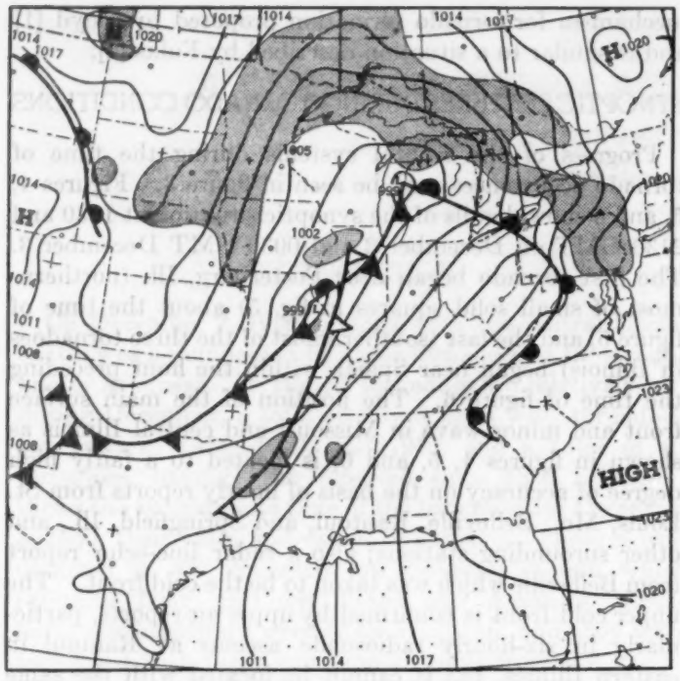

FIGURE 4.—Surface weather chart for 1830 GMT, December 2, 1950.

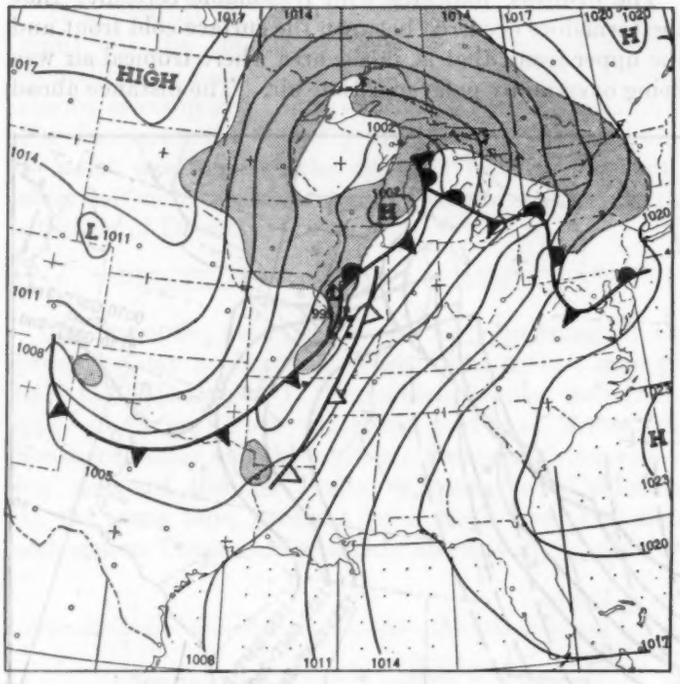

FIGURE 5.—Surface weather chart for 2130 GMT, December 2, 1950.

of the cold front is estimated in each case to be of the order of 15 or 20 miles, based on interpolated positions of the cold front; the cold front appears to have arrived about thirty minutes after beginning of the tornado. The tornadoes in Illinois were south of the center or apex of a small wave along the cold front, the first tornado estimated to be about 40 miles south-southeast of the moving center,

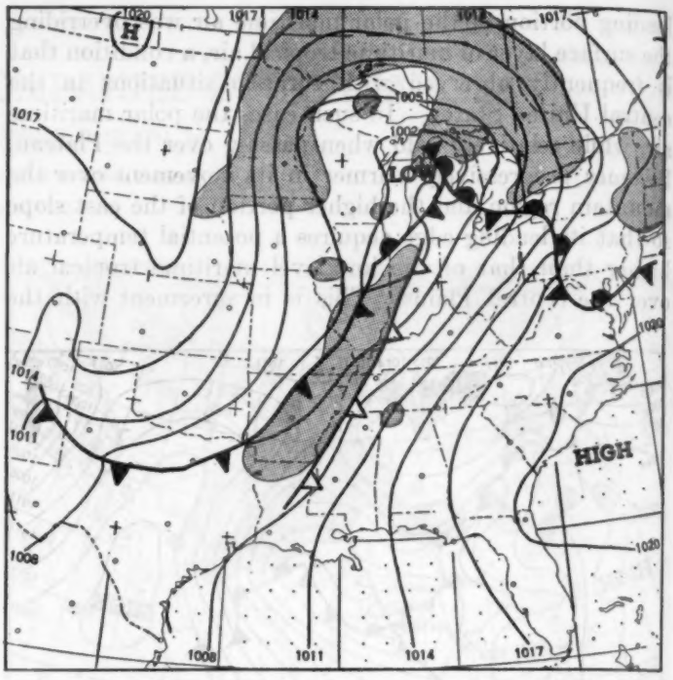

FIGURE 6.—Surface weather chart for 0030 GMT, December 3, 1950.

the second about 90 miles directly south, and the third about 180 miles south-southwest. Taken in the order of time of occurrence, each of the Illinois tornadoes was progressively farther southward and farther from the apex of the wave. It is possible that the Arkansas tornado also occurred in connection with a very minor wave along the front, as is suggested in figure 6 by the shape of the cold front in that area, but evidence for the wave is uncertain.

Occurrence of the tornadoes behind rather than along the upper cold front suggests that the necessary degree of vertical instability did not develop upon arrival of the leading edge of the colder air aloft, but that instability became sufficient as the polar maritime air aloft became deeper and colder.

Significant features of the upper air structure may be seen in figures 7–10, inclusive. Figures 7 and 8, containing radiosonde data from Little Rock, Ark., and Rantoul, Ill., respectively, illustrate the temperature and moisture distribution in the warm air. In figure 7, for Little Rock, the sounding for 1500 GMT on the 2d (before arrival of the polar maritime air aloft) shows moist convectively unstable air up to 750 mb. (about 8,000 ft. msl), capped by an inversion above which the relative humidity was noticeably lower. The zero (°C) wet-bulb temperature was at 700 mb., immediately above the inversion, and the air from the top of the inversion up to 560 mb. was also convectively unstable. The second sounding in figure 7, for 0300 GMT on the 3d, was taken about the time of arrival of the surface cold front, yet cooling had occurred between 750 and 600 mb., wiping out the inversion and indicating arrival of the polar maritime air aloft; this is further indicated by higher relative humidity above 750 mb.

The Rantoul sounding for 1500 GMT, December 2 (fig. 8) is similar to the Little Rock sounding for the same time. The 2100 GMT sounding shows slight cooling between 780 and 580 mb. and a noticeable increase of moisture above 700 mb. The sounding for 0300 GMT of the 3d shows both the arrival of the surface cold front and further cooling at higher levels associated with the cold front aloft.

Figures 9 and 10 are vertical cross sections for 1500 GMT of the 2d and 0300 GMT of the 3d, respectively, taken approximately through the tornado area. The inversion or stable layer, which is shown between 6000 and 8000 feet at Rantoul in figure 9, was capping the lower warm moist air and was accompanied by much drier air above. This is typical of conditions preceding tornadoes. Winds of around 50 knots within the lower layer of warm moist air, as may be seen at St. Louis and Springfield in figure 9, are also typical of wind conditions frequently observed immediately ahead of tornadoes. Whether or not there was marked cyclonic shear along the line of tornado formation cannot be determined from available data, but strong southerly flow ahead of the surface front suggests this possibility; such a line of shear (or local zone of cyclonic rotation) is necessary at least in the immediate vicinity of a tornado if we accept as a requirement for tornado formation that there must be low-level convergence within a zone of marked cyclonic vorticity.

Figure 10 shows the eastward extension of the cold front aloft with respect to the surface cold front and the consequent steepening of the lapse rate immediately above the lower moist air. It is not intended, however, to suggest that all steepening of the lapse rate aloft ahead of the surface cold front was the result of intrusion aloft of a colder air mass. While the intrusion of colder air seems clearly indicated in this case, as well as with tornado situations in general, there is reason to believe that vertical motion aloft over a surface warm sector may also be a factor tending to increase the lapse rate.

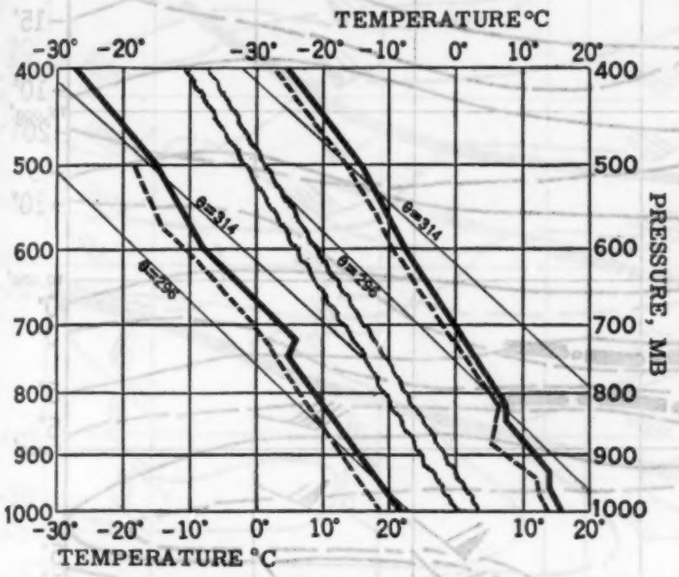

FIGURE 7.—Radiosonde observations at Little Rock, Ark., plotted on a pseudo-adiabatic diagram: Left—1500 GMT, December 2, 1950; Right—0300 GMT, December 3, 1950. Dashed curves are wet-bulb temperature.

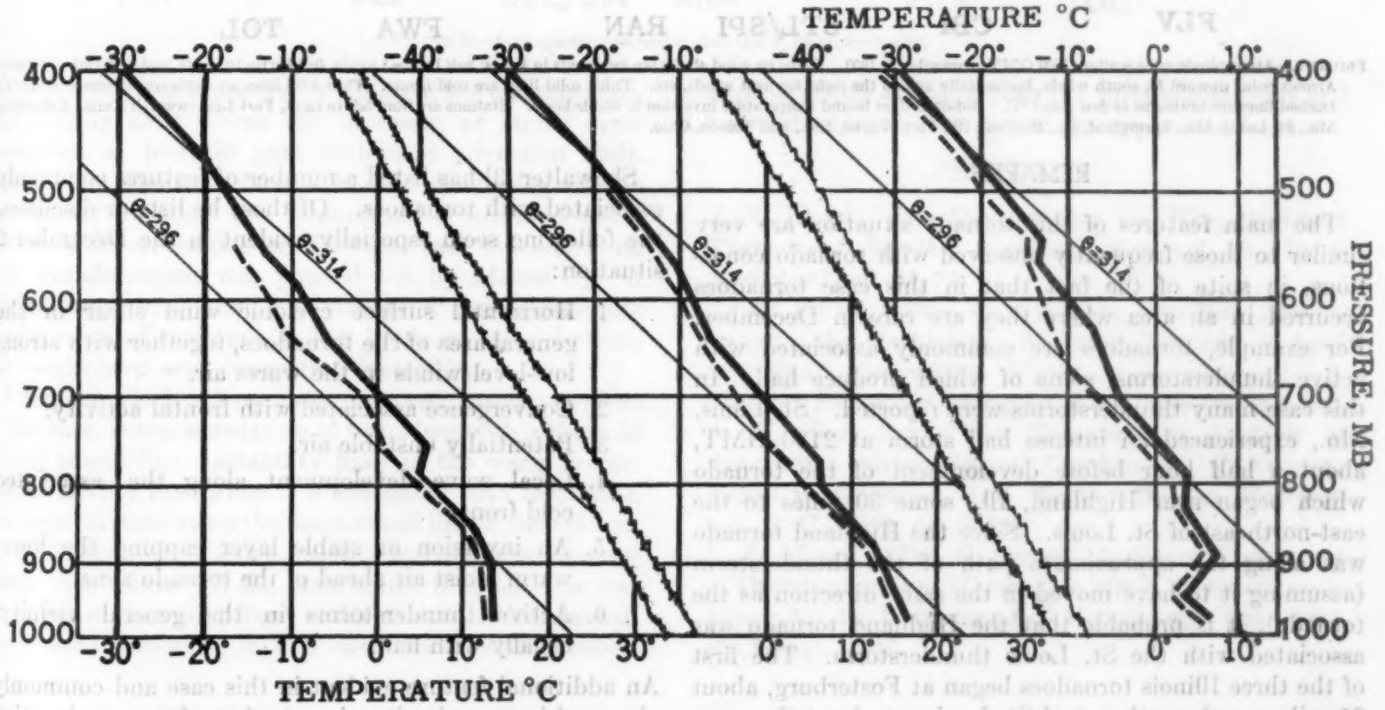

FIGURE 8.—Radiosonde observations at Rantoul, Ill., plotted on a pseudo-adiabatic diagram: Left—1500 GMT, December 2, 1950; Center—2100 GMT, December 2, 1950; Right—0300 GMT, December 3, 1950. Dashed curves are wet-bulb temperature.

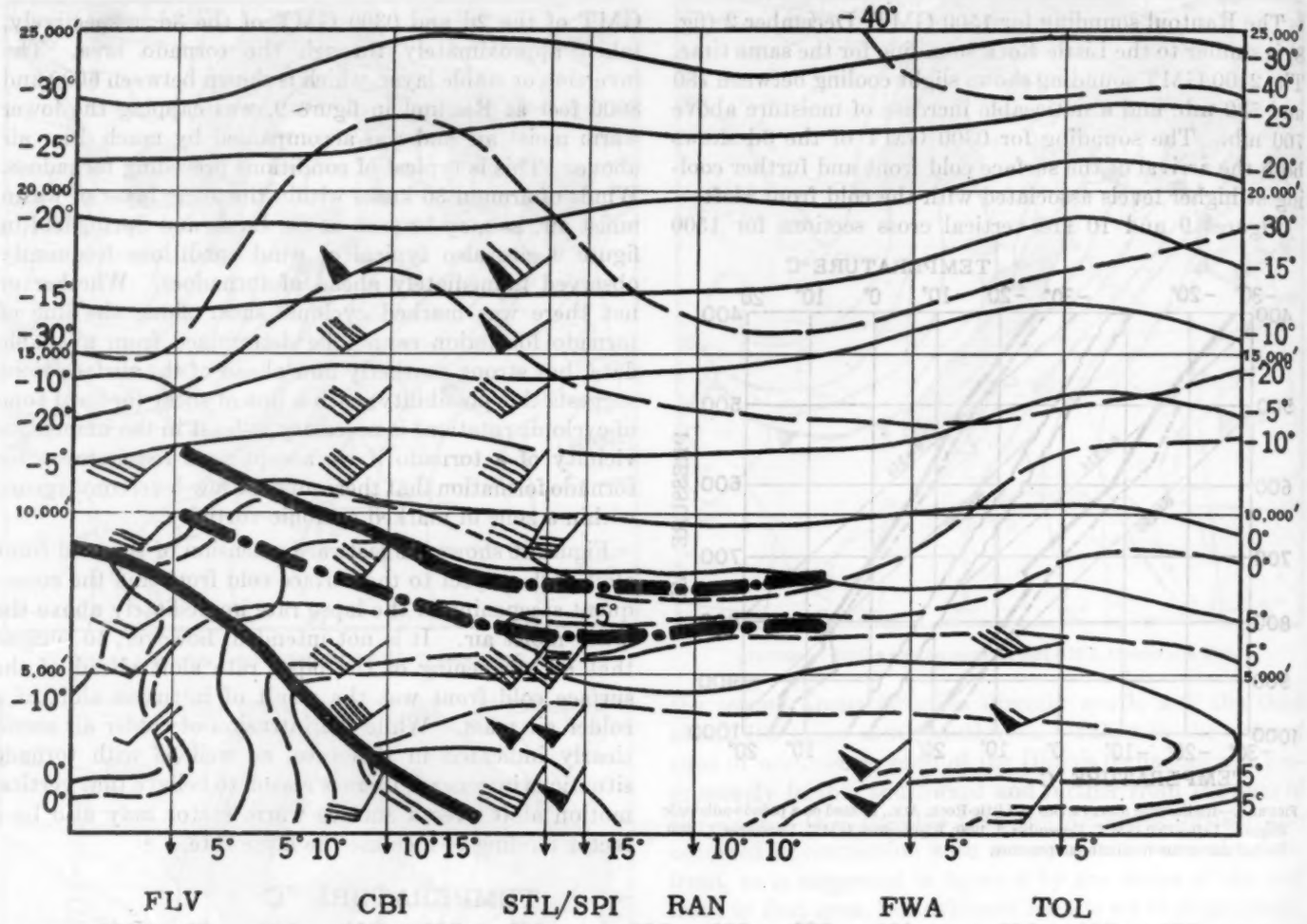

FIGURE 9.—Atmospheric cross section, 1500 GMT, December 2, 1950. Barbs on wind shafts are for speeds in knots; half barb=5 knots, full barb=10 knots, and pennant=50 knots. Arrows point upward for south winds, horizontally and to the right for west winds, etc. Thick solid lines are cold fronts. Thin solid lines are isotherms of temperature °C. Dashed lines are isotherms of dew point °C. Dot-dash lines bound temperature inversion or stable layer. Stations are from left to right, Fort Leavenworth Kans., Columbia, Mo., St. Louis, Mo., Springfield, Ill., Rantoul, Ill., Fort Wayne, Ind., and Toledo, Ohio.

REMARKS

The main features of this tornado situation are very similar to those frequently observed with tornado conditions, in spite of the fact that in this case tornadoes occurred in an area where they are rare in December. For example, tornadoes are commonly associated with active thunderstorms, some of which produce hail. In this case many thunderstorms were reported. St. Louis, Mo., experienced an intense hail storm at 2130 GMT, about a half hour before development of the tornado which began near Highland, Ill., some 30 miles to the east-northeast of St. Louis. Since the Highland tornado was along the approximate path of the thunderstorm (assuming it to have moved in the same direction as the tornado), it is probable that the Highland tornado was associated with the St. Louis thunderstorm. The first of the three Illinois tornadoes began at Fosterburg, about 25 miles north-northeast of St. Louis, at about the same time that the thunderstorm was occurring at St Louis.

Showalter [3] has listed a number of features commonly associated with tornadoes. Of those he lists or discusses, the following seem especially evident in the December 2 situation:

1. Horizontal surface cyclonic wind shear in the general area of the tornadoes, together with strong low-level winds in the warm air.
2. Convergence associated with frontal activity.
3. Potentially unstable air.
4. Local wave development along the associated cold front.
5. An inversion or stable layer capping the lower warm moist air ahead of the tornado zone.
6. Active thunderstorms in the general vicinity, usually with hail.

An additional feature evident in this case and commonly observed in tornado situations is that of warm advection in the low levels (especially the lower 2,000 or 3,000 feet

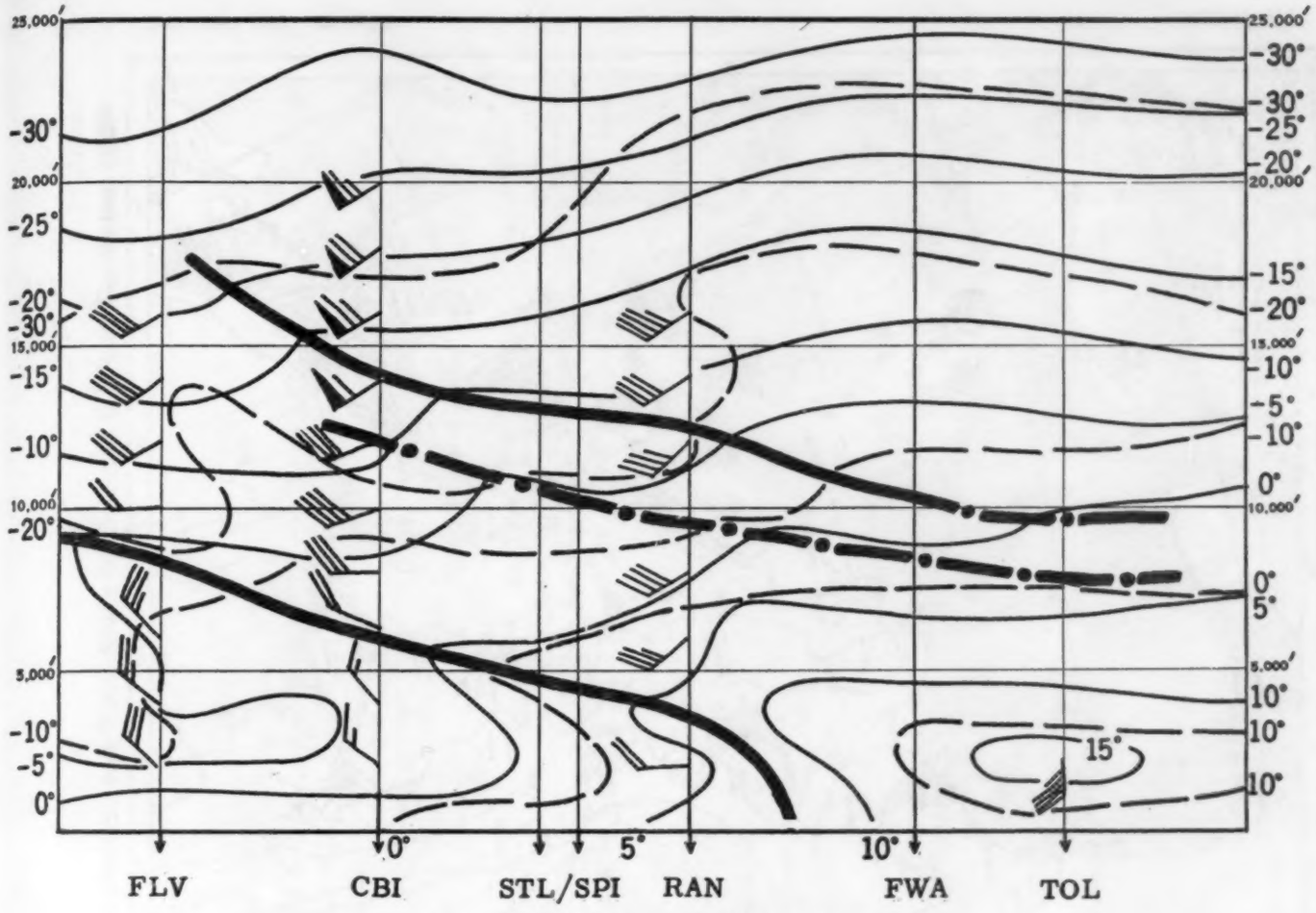

FIGURE 10.—Atmospheric cross section, 0300 GMT, December 3, 1950.

above the ground) immediately ahead of the tornado zone, and cooling aloft above the inversion or stable layer associated at least in part with cold advection aloft. These are important factors in the steepening of the vertical lapse rate. The importance of low-level warm advection in steepening of the lapse rate in connection with thunderstorms was pointed out by Means [4]. A related factor in this and with tornado situations generally is advection of moisture at low levels so as to increase the conditional and convective instability.

A feature which is frequently observed, but not evident in this case, is the association of tornadoes with a pre-cold frontal squall line (instability line) in the warm sector. The instability line, when it is associated with tornadoes, is at least in most cases the main squall line of the system. In the December 2 situation, the tornadoes were associated with (ahead of) the main squall line, but this was along the surface cold front so far as can be determined. There were, however, in this case, pre-frontal thunderstorms occurring ahead of both the surface cold front and the tornadoes but to the rear of the upper cold front.

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4. L. L. Means, "The Nocturnal Maximum Occurrence of Thunderstorms in Midwestern States," *Miscellaneous Report* No. 16, Department of Meteorology, University of Chicago, 1945.

U. S. GOVERNMENT PRINTING OFFICE: 1951

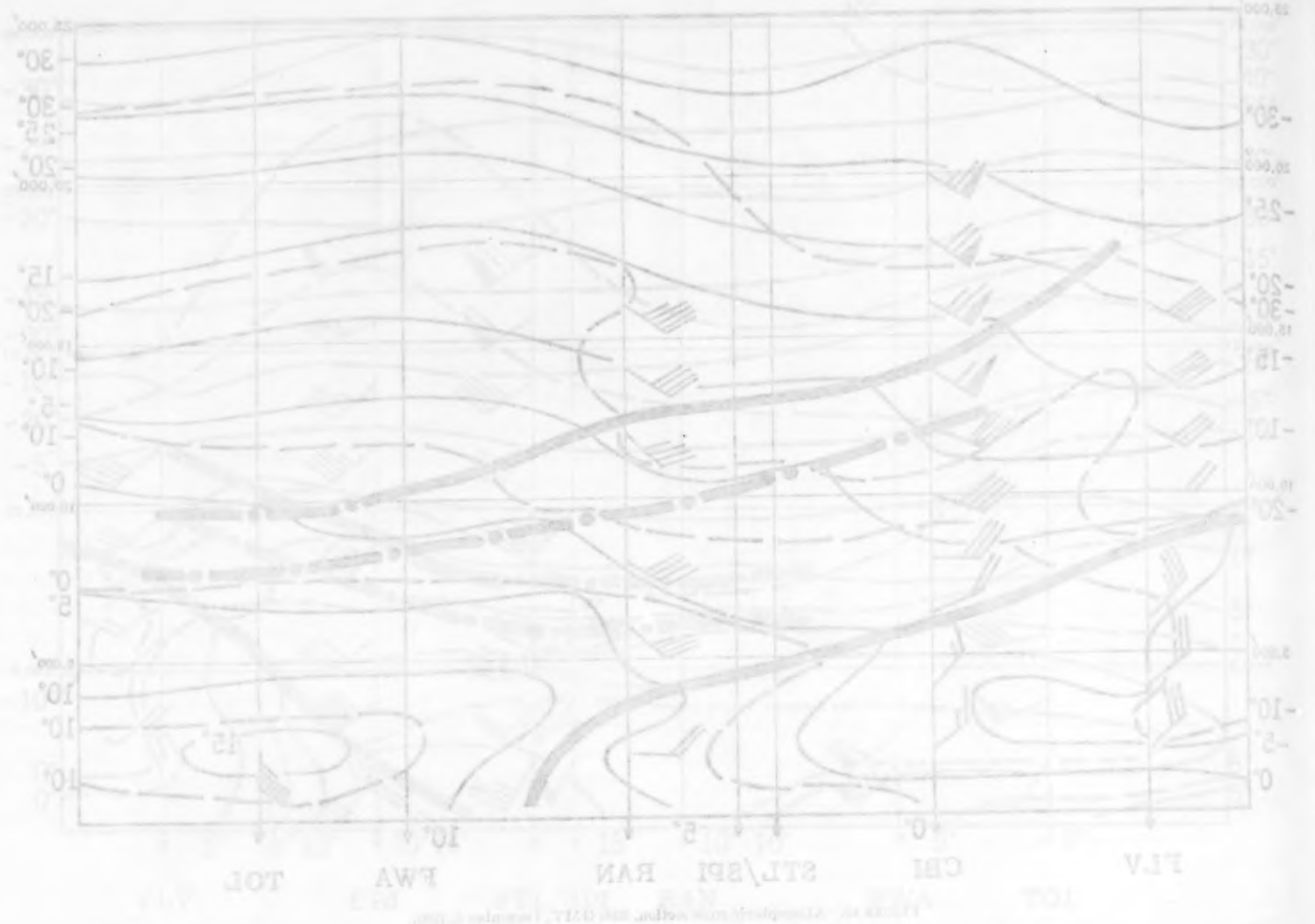
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Chart I. Departure (°F.) of the Mean Temperature from the Normal, and Wind Roses for Selected Stations, December 1950

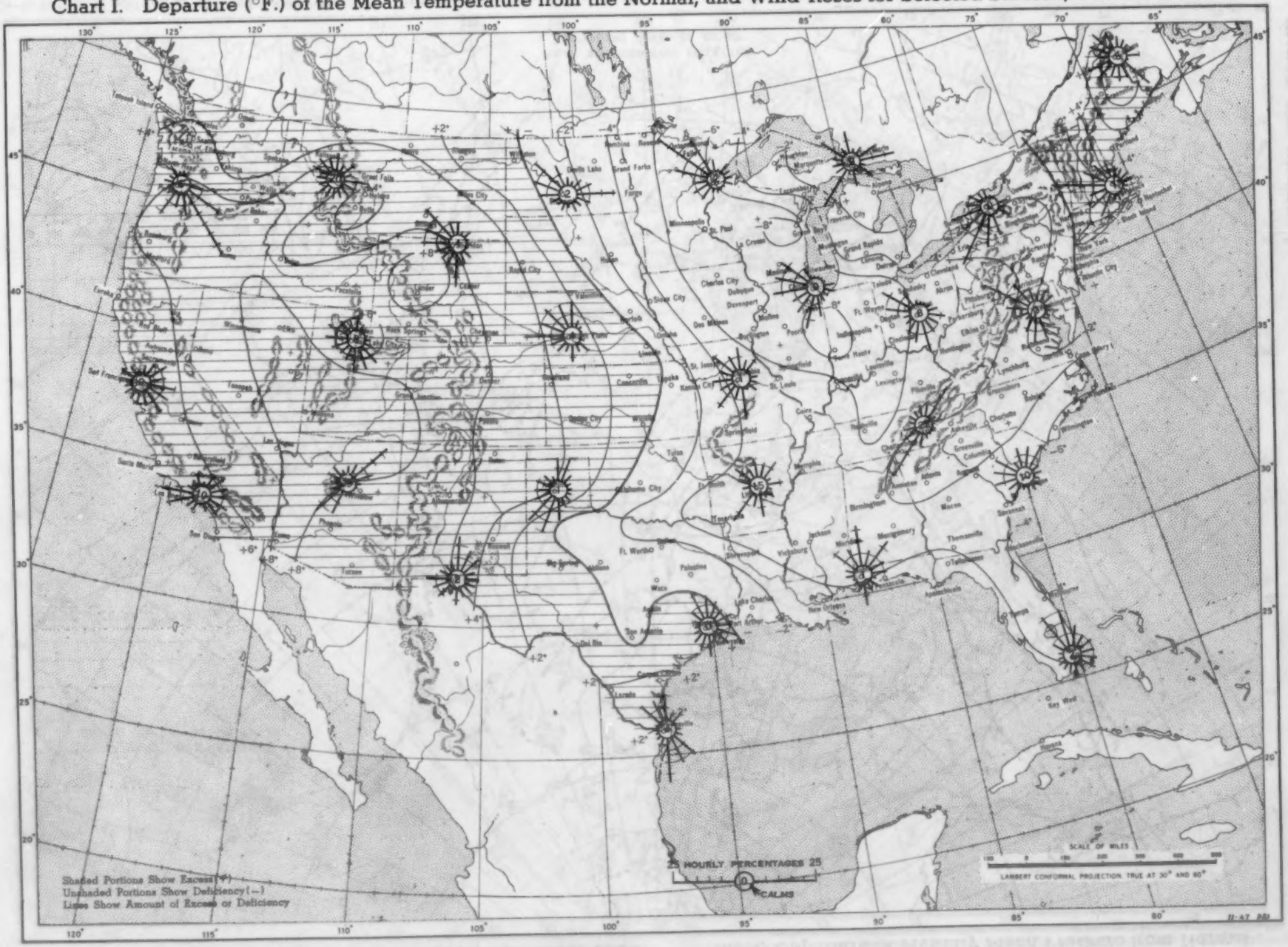
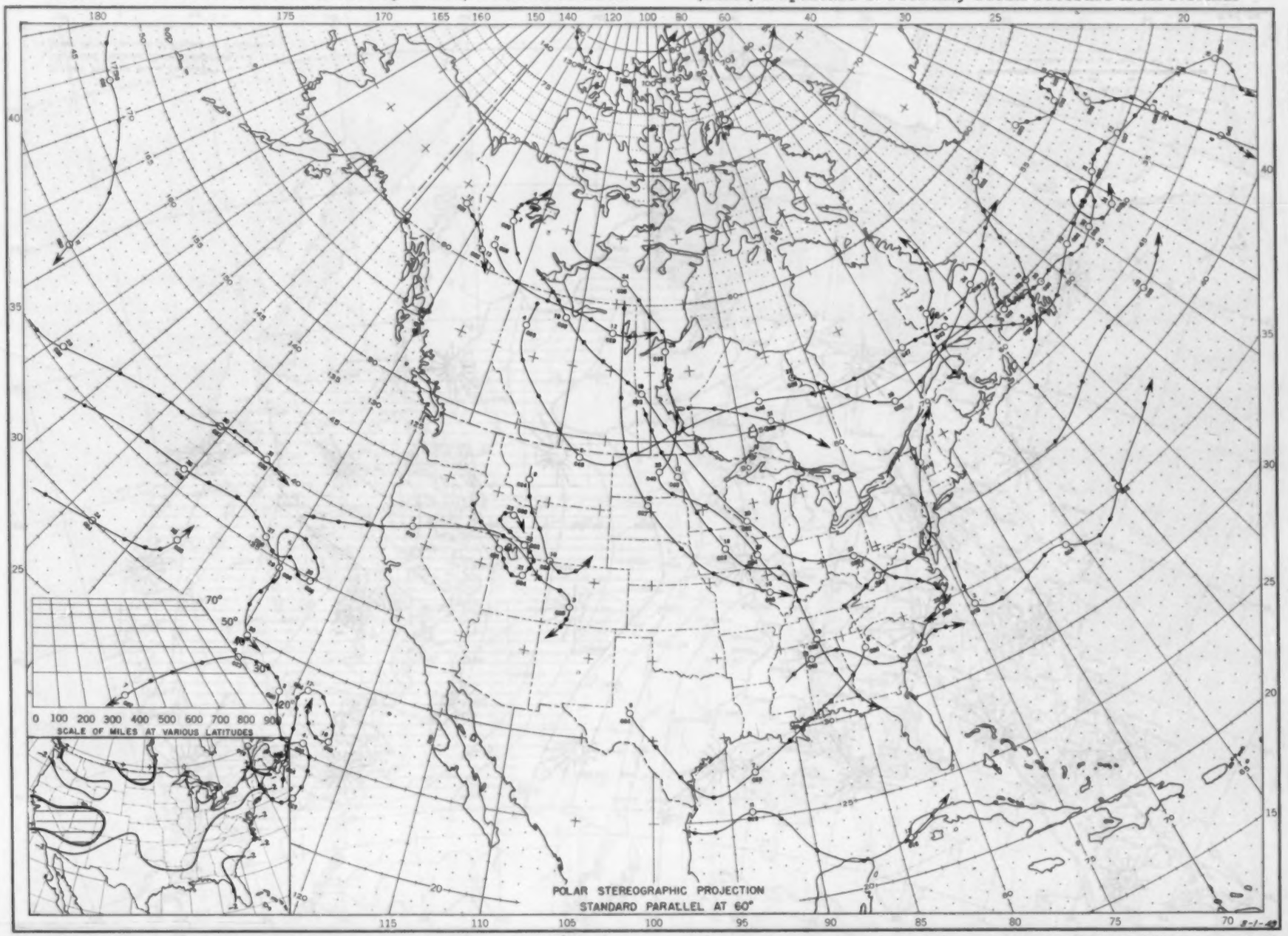
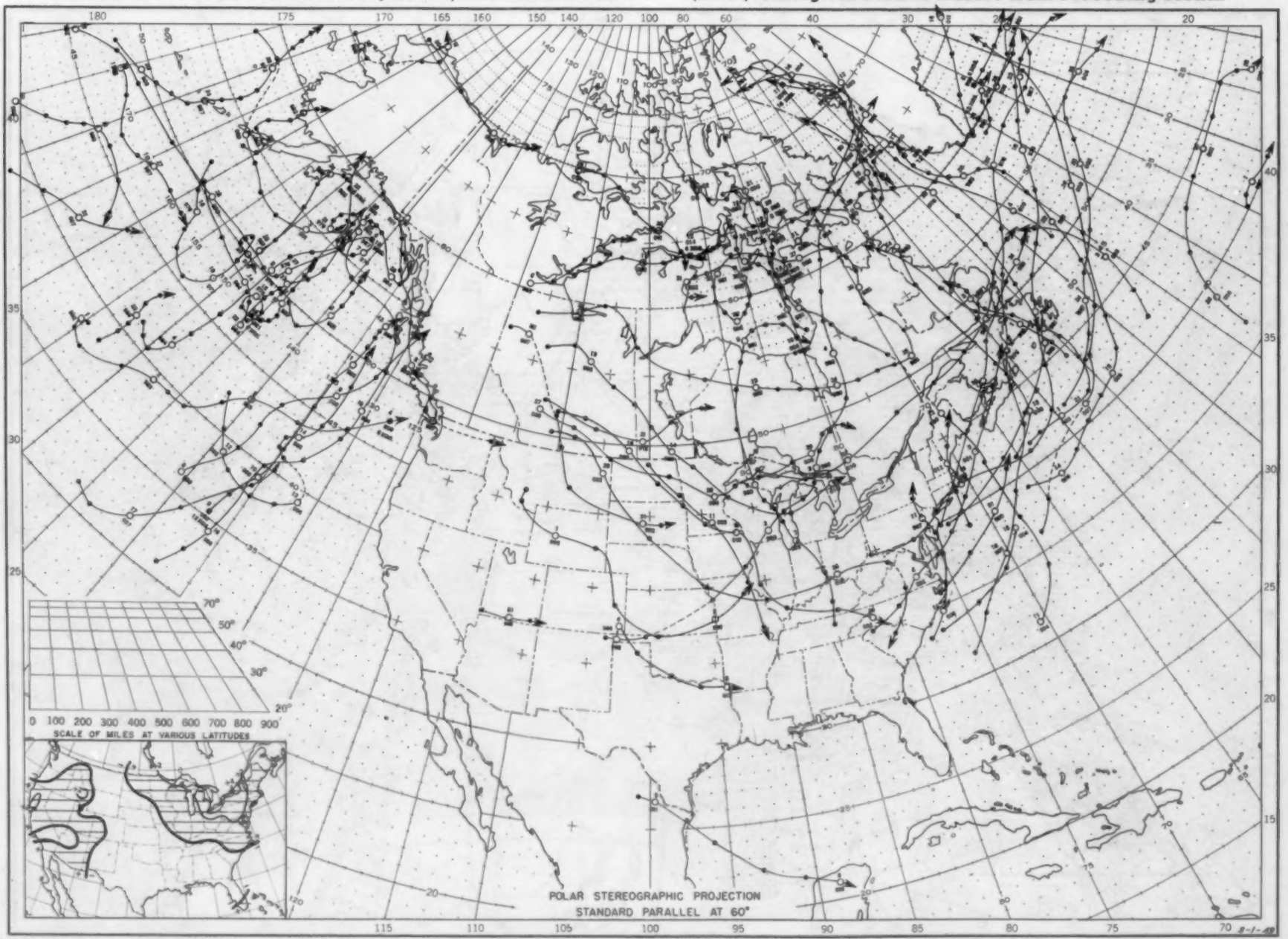

Shaded Portions Show Excess (+)
Unshaded Portions Show Deficiency (−)
Lines Show Amount of Excess or Deficiency
25 HOURLY PERCENTAGES 25
CALMS
SCALE OF MILES
LAMBERT CONFORMAL PROJECTION TRUE AT 30° AND 60°

Chart II. Tracks of Centers of Anticyclones, December 1950. (Inset) Departure of Monthly Mean Pressure from Normal



Circle indicates position of anticyclone at 7:30 a. m. (75th meridian time). Figure above circle indicates date, and figure below, pressure to nearest millibar. Dots indicate intervening 6-hourly positions. Square indicates position of stationary center for period shown. Only those centers which could be identified for 24 hours or more are included.

Chart III. Tracks of Centers of Cyclones, December 1950. (Inset) Change in Mean Pressure from Preceding Month



Circle indicates position of cyclone at 7:30 a. m. (75th meridian time). Figure above circle indicates date, and figure below, pressure to nearest millibar. Dots indicate intervening 6-hourly positions. Square indicates position of stationary center for period shown. Only those centers which could be identified for 24 hours or more are included.

Chart IV. Percentage of Clear Sky Between Sunrise and Sunset, December 1950

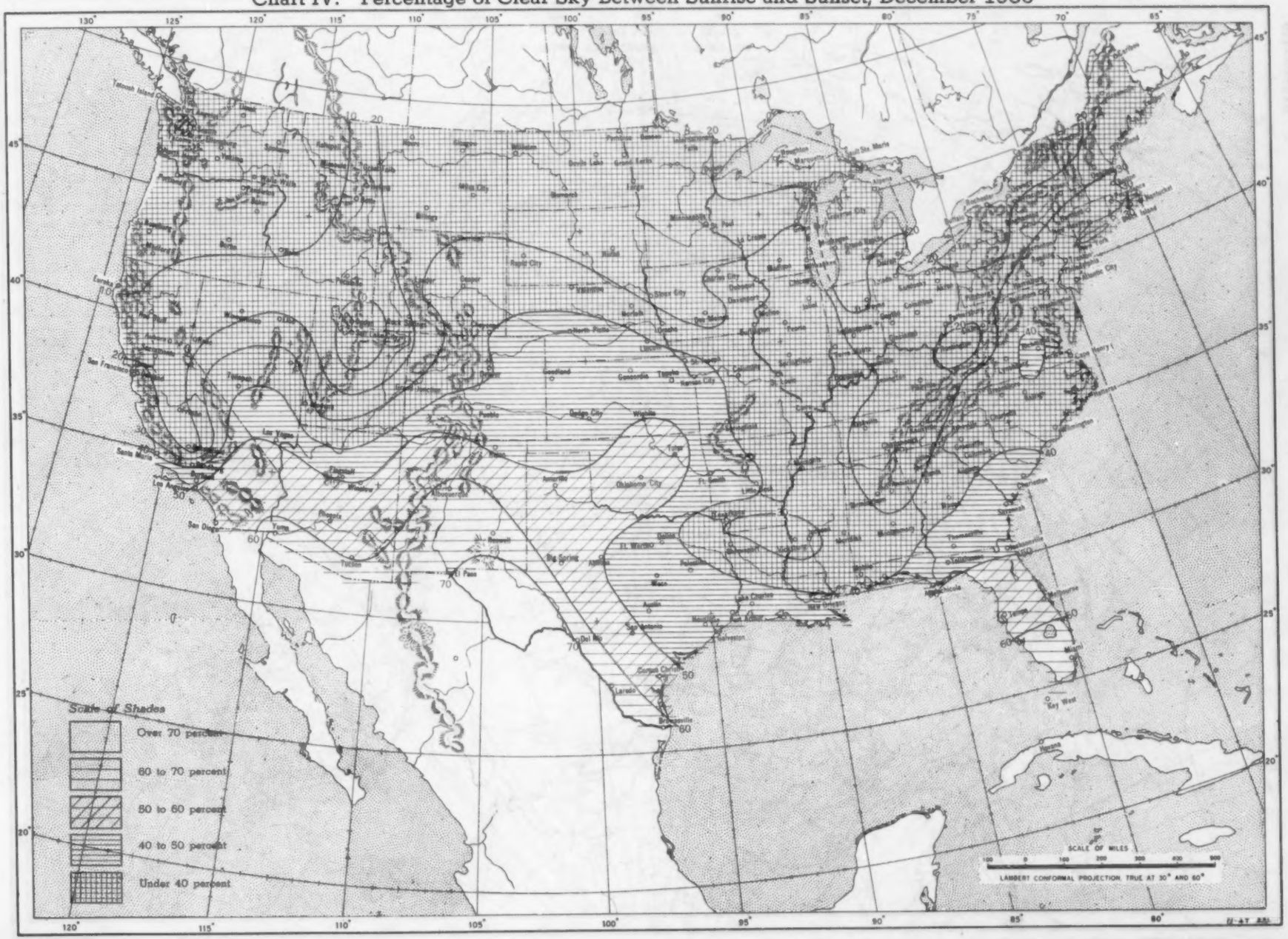

Scale of Shades
Over 70 percent
60 to 70 percent
50 to 60 percent
40 to 50 percent
Under 40 percent
SCALE OF MILES
LAMBERT CONFORMAL PROJECTION TRUE AT 30° AND 60°

Chart V. Total Precipitation, Inches, December 1950. (Inset) Departure of Precipitation from Normal

Scale of Shades
0 to 1 Inch
1 to 2 Inches
2 to 4 Inches
4 to 6 Inches
Over 6 Inches
LAMBERT CONFORMAL PROJECTION TRUE AT 30° AND 60°
SCALE OF MILES

Chart VI. Mean Isobars (mb.) at Sea Level and Mean Isotherms (°F.) at Surface., December 1950

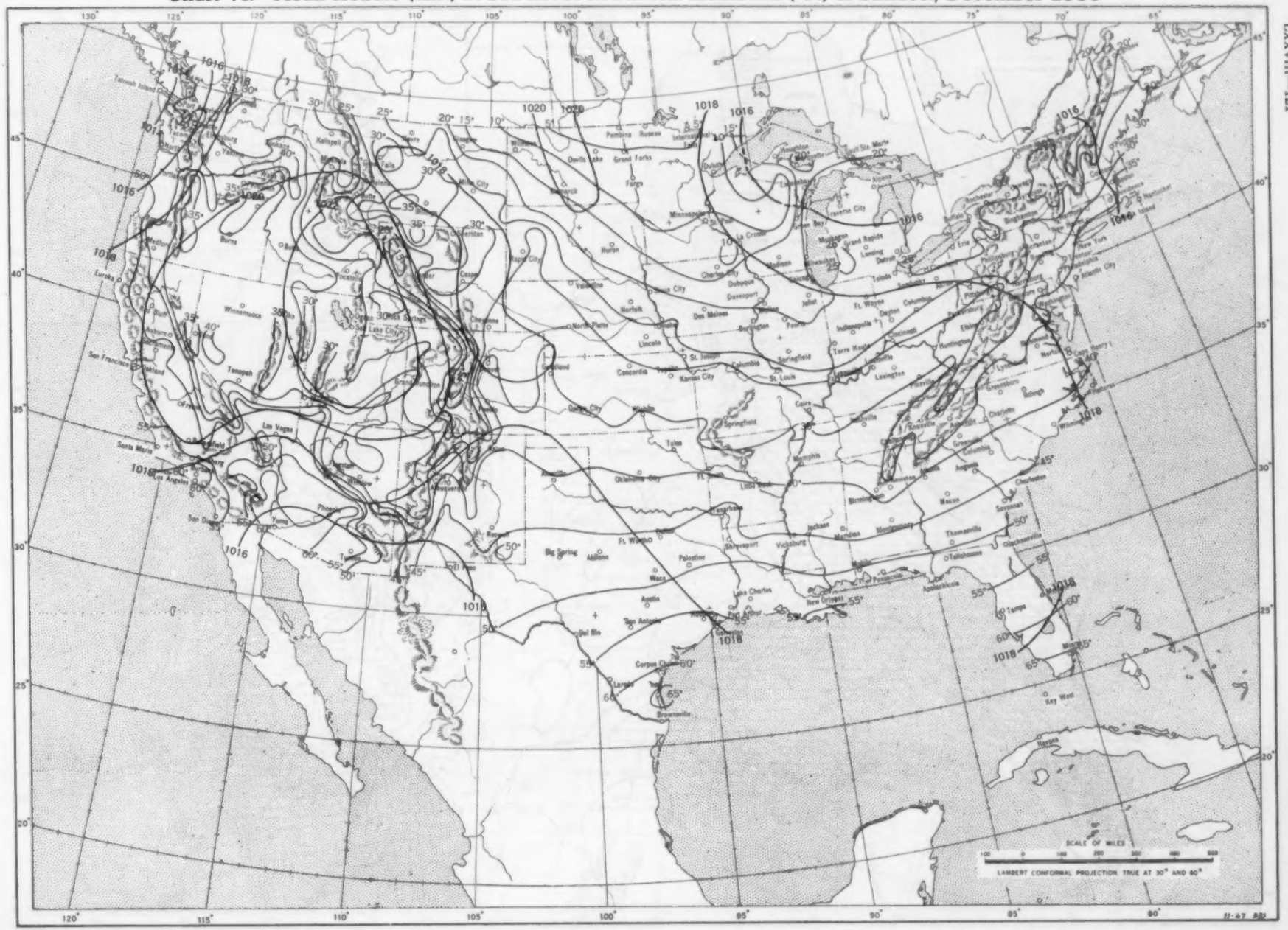
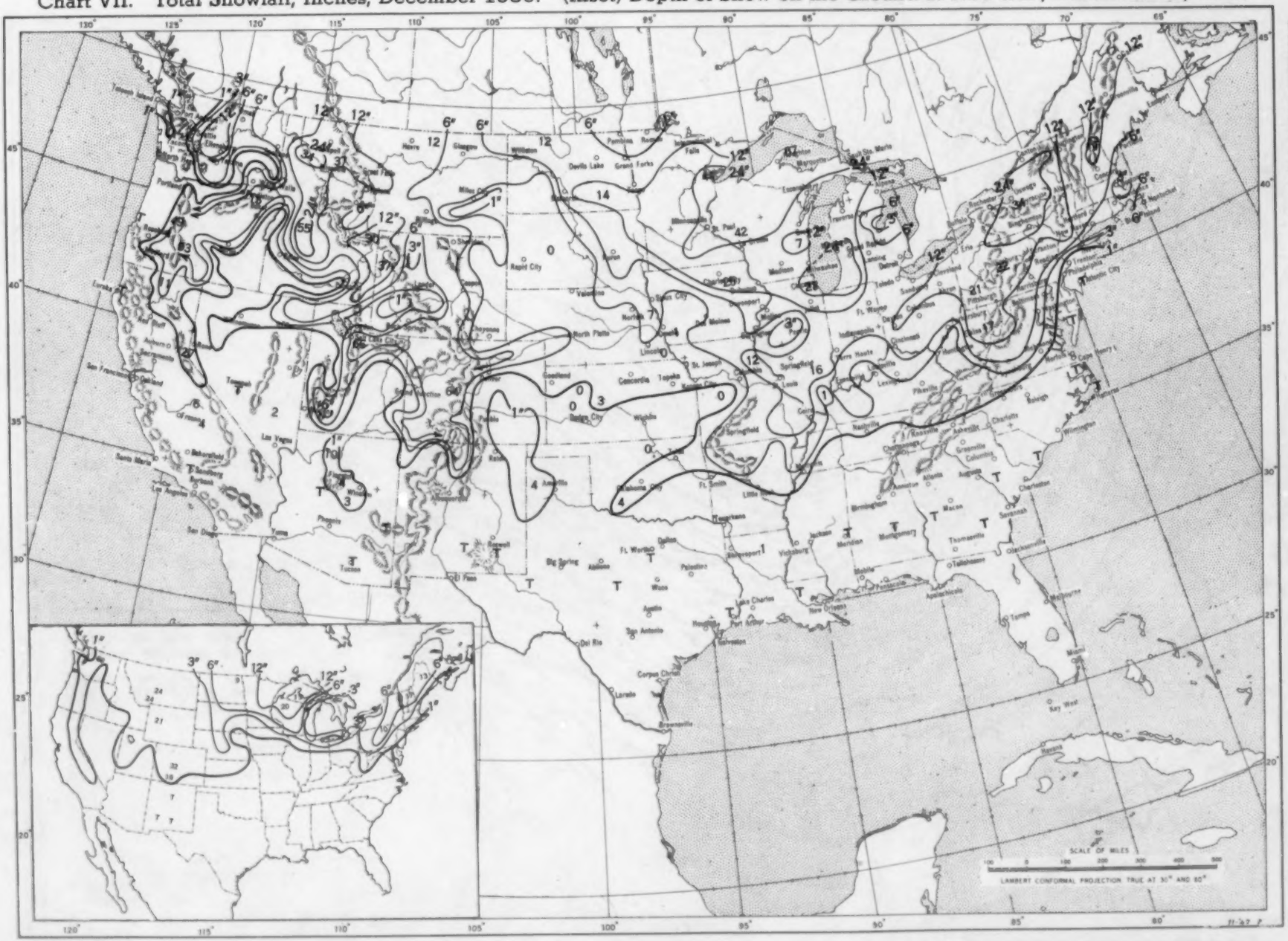


Chart VII. Total Snowfall, Inches, December 1950. (Inset) Depth of Snow on the Ground at 7:30 a. m., December 26, 1950



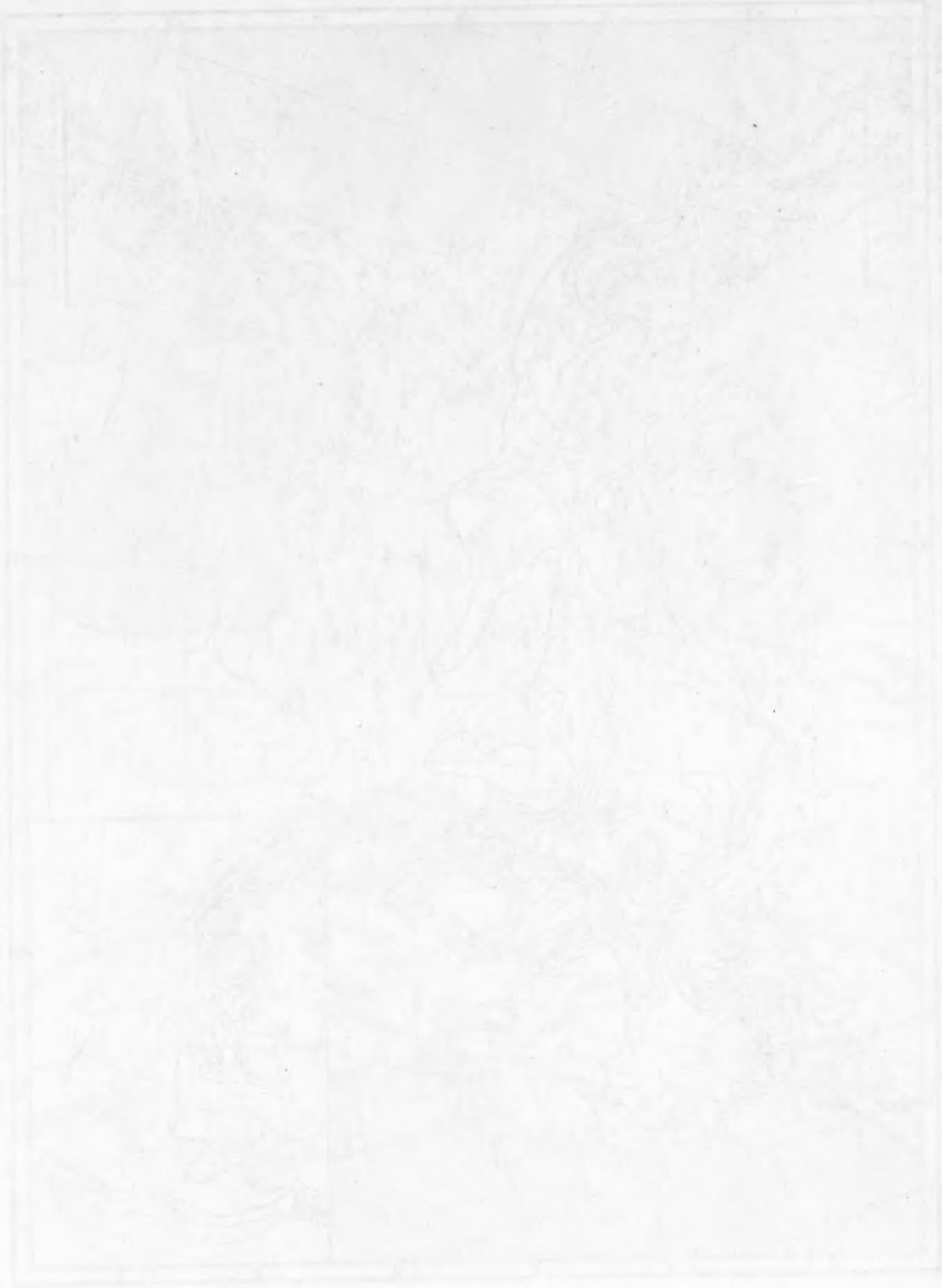
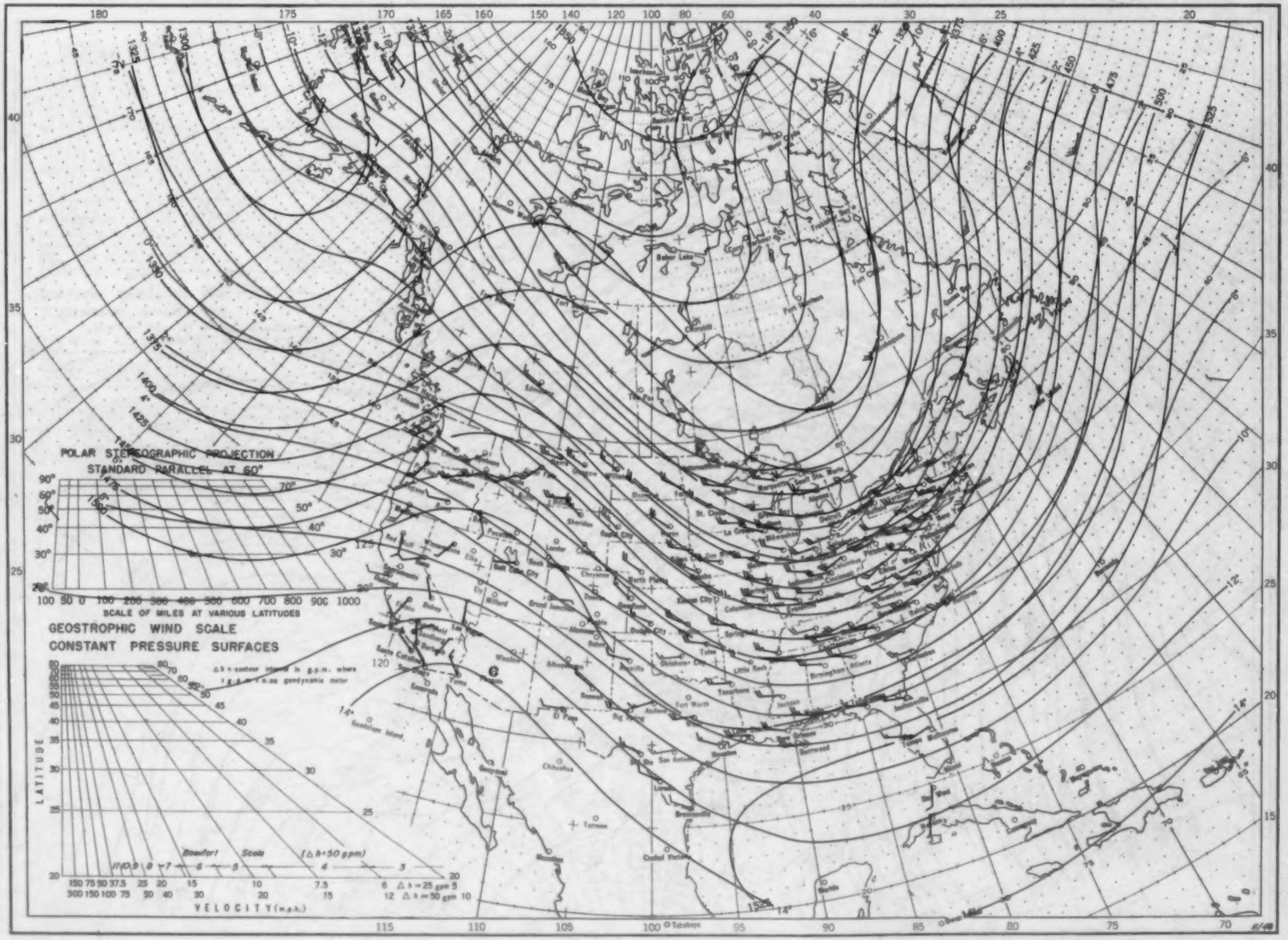
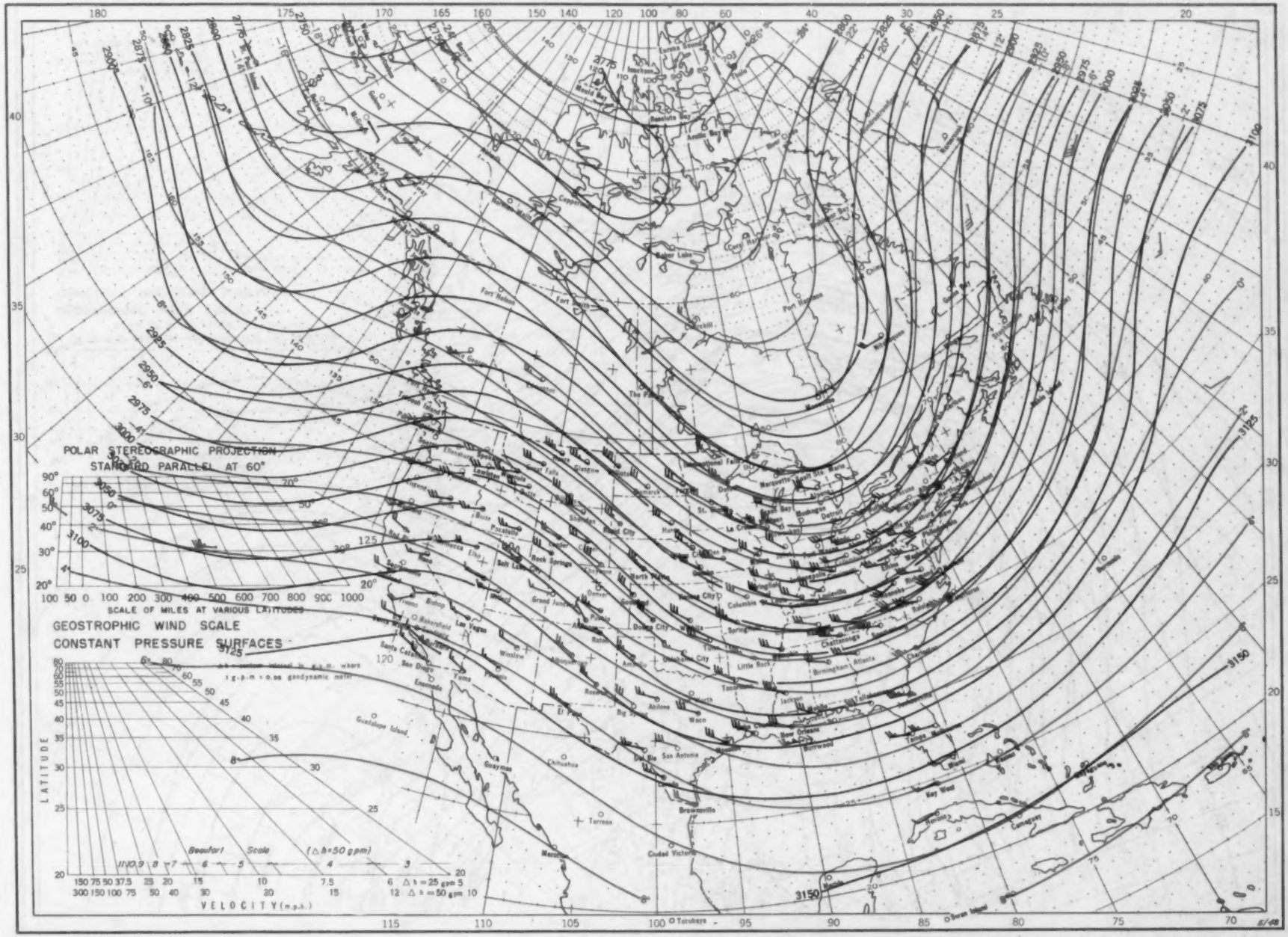


Chart VIII, December 1950. Contour Lines of Mean Dynamic Height (Geopotential) in Units of 0.98 Dynamic Meters and Mean Isotherms in Degrees Centigrade for the 850-millibar Pressure Surface, and Resultant Winds at 1,500 Meters (m. s. l.)



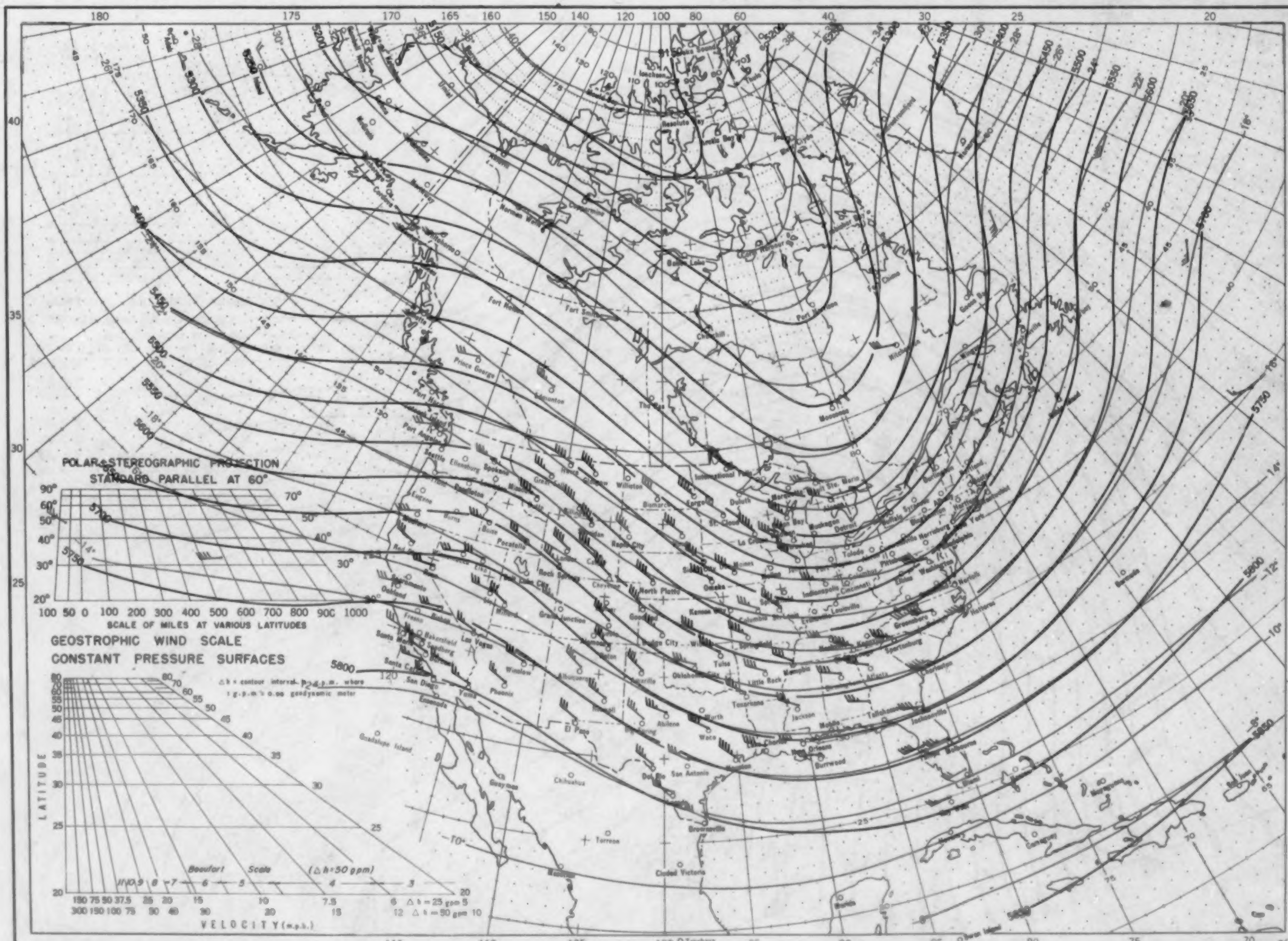
Contour lines and isotherms based on radiosonde observations at 0300 G. C. T. Winds indicated by black arrows based on pilot balloon observations at 2100 G. C. T.; those indicated by red arrows based on rawins taken at 0300 G. C. T.

Chart IX, December 1950. Contour Lines of Mean Dynamic Height (Geopotential) in Units of 0.98 Dynamic Meters and Mean Isotherms in Degrees Centigrade for the 700-millibar Pressure Surface, and Resultant Winds at 3,000 Meters (m. s. l.)



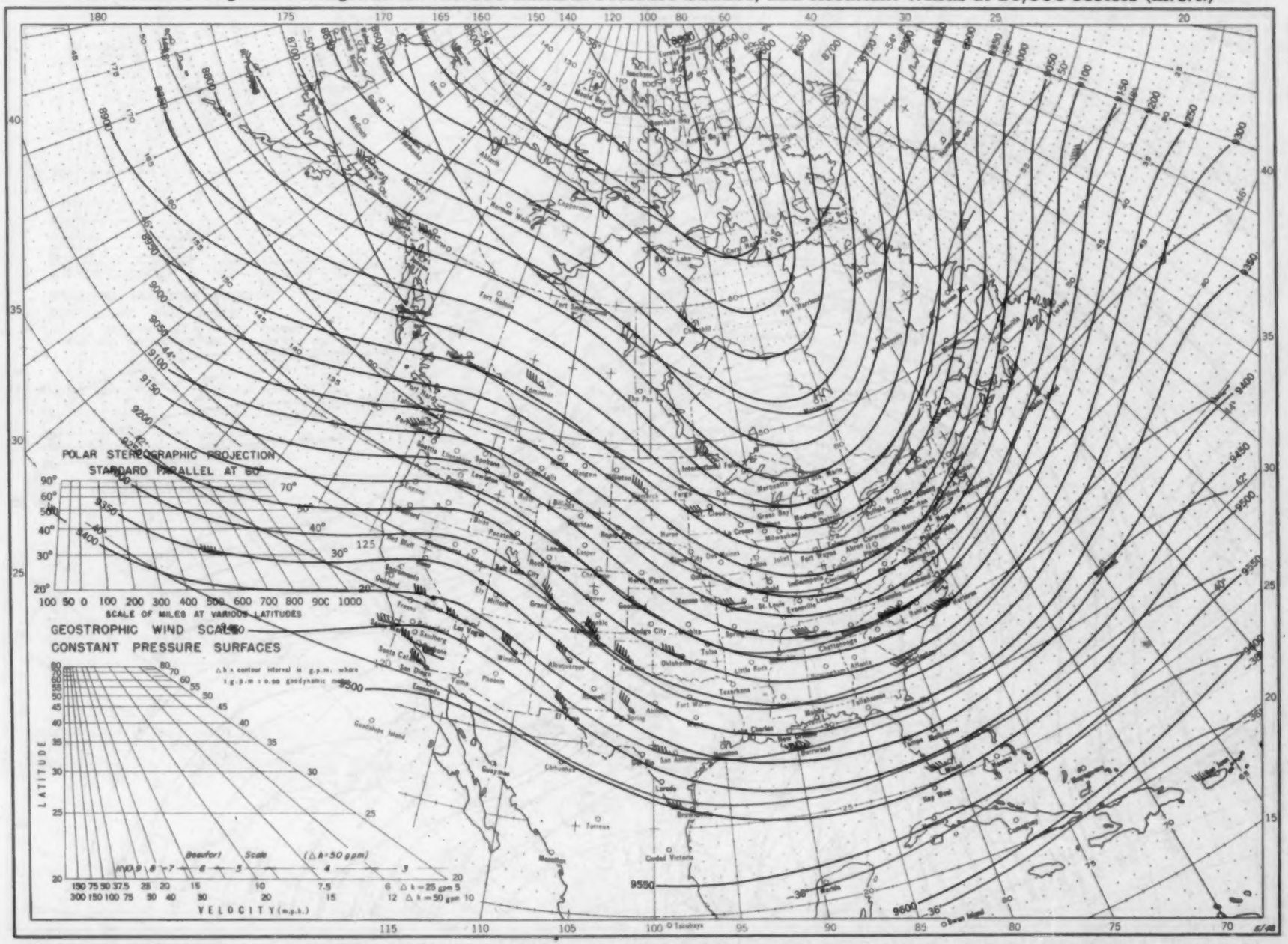
Contour lines and isotherms based on radiosonde observations at 0300 G. C. T. Winds indicated by black arrows based on pilot balloon observations at 2100 G. C. T.; those indicated by red arrows based on rawins taken at 0300 G. C. T.

Chart X, December 1950. Contour Lines of Mean Dynamic Height (Geopotential) in Units of 0.98 Dynamic Meters and Mean Isotherms in Degrees Centigrade for the 500-millibar Pressure Surface, and Resultant Winds at 5,000 Meters (m. s. l.)



Contour lines and isotherms based on radiosonde observations at 0300 G. C. T. Winds indicated by black arrows based on pilot balloon observations at 2100 G. C. T.; those indicated by red arrows based on rawins taken at 0300 G. C. T.

Chart XI, December 1950. Contour Lines of Mean Dynamic Height (Geopotential) in Units of 0.98 Dynamic Meters and Mean Isotherms in Degrees Centigrade for the 300-millibar Pressure Surface, and Resultant Winds at 10,000 Meters (m. s. l.)



Contour lines and isotherms based on radiosonde observations at 0300 G. C. T. Winds indicated by black arrows based on pilot balloon observations at 2100 G. C. T.; those indicated by red arrows based on rawins taken at 0300 G. C. T.

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